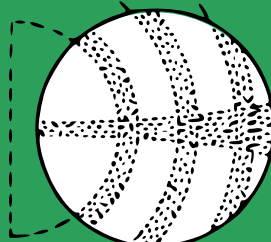
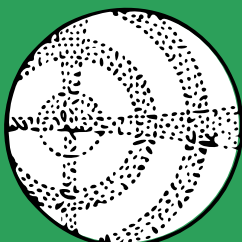
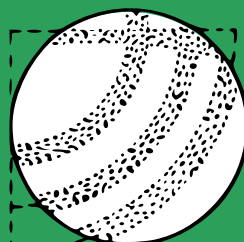
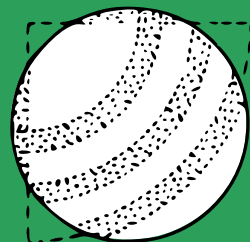
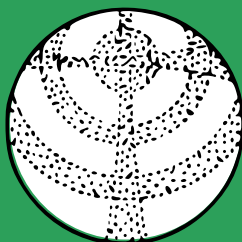
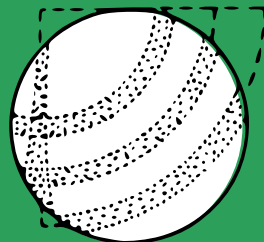
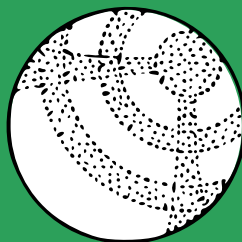
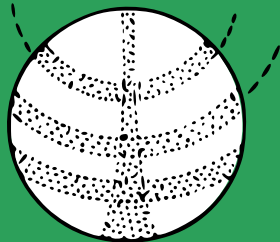
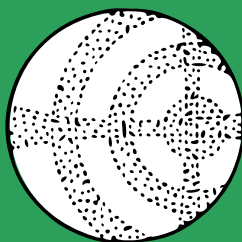


V. Zubkov

GENERAL PETROGRAPHY

A Short Course

Mir
Publishers
Moscow



В. Зубков

**КРАТКИЙ КУРС
ОБЩЕЙ ПЕТРОГРАФИИ**

**ИЗДАТЕЛЬСТВО «НЕДРА»
МОСКВА**

На английском языке

V. ZUBKOV

General Petrography

A SHORT COURSE

MIR
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CHAPTER I

Concepts concerning petrography and rocks

Petrography is the study of rocks. The present tendency is to use the term petrography only for the description of rocks, their compositions and structures, and to give the name *petrology* to all questions connected with their genesis.

The science of sedimentary rocks is separated into a special science, *lithology*. There are also branches of petrography under the names of experimental and technical petrography and petrochemistry.

Experimental petrography is concerned with the formation processes of rocks under experimental conditions in the laboratory. In a number of cases laboratory experiment is the only means of resolving disputed problems. A particular example is the experimental demonstration that plagioclases form a continuous series of solid solutions of albite and anorthite, and not definite combinations of these two components in the form of independent minerals (oligoclase, andesine, labradorite).

Technical petrography, which has been created recently by the work of Academician D. Belyankin and his coworkers and followers, aims at the study of artificial stones by petrographical methods, and offers advice for their preparation wherever they are obtained in the course of metallurgical, ceramic and other technical processes.

Petrochemistry, the investigator of which was Academician A. Zavaritskii, helps to reveal the fundamental laws of the chemical composition of the different groups of igneous rocks, and their changes during metamorphism.

Rocks are mineral aggregates made up of one or more minerals, and occupying a considerable part of the earth's crust. They are characterized by a greater or lesser constancy of chemical and mineralogical composition and definite conditions of occurrence. Petrography is thus closely connected with all geological disciplines, especially with mineralogy, as well as with physical chemistry.

Rocks are divided into three main genetic groups, depending on the geological processes which formed them: magmatic (or igneous), sedimentary and metamorphic.

Magmatism manifests itself in the earth's crust in two main forms. Firstly, the magma as a silicate melt from the depths of the earth can pour out on to the surface through an open crack in the crust (volcanoes) in the form of lava, and cool rapidly, so giving rise to the effusive rocks. Secondly, the magma may not reach the surface, and a deep-seated (intrusive) rock is then formed as a result of slow cooling at depth.

Sedimentary rocks are formed of materials which are the result of the disintegration of igneous rocks by weathering to form either fragments visible to the naked or aided eye, or colloidal bodies and solutions. All this is carried away by running water into low lying areas, or into oceans and lakes, where it is deposited either by the drying up of the water, or by settling on the bottom as a sediment; sea waves, glaciers and wind also play their part in this transfer and deposition.

Mineral fragments undergo mechanical sorting during denudation and are thus separated according to size and specific gravity. In particular, ores and other valuable minerals, gold, platinum, cassiterite, diamonds and similar so-called placer deposits of useful minerals are formed in this way.

The separation of colloidal and dissolved substances takes place in such a way that the various colloids and solutes are precipitated as a sediment at various times and under various conditions. For example, the iron oxide compounds are thus precipitated quickly, while common salt remains in solution for a very long time. In one case, the deciding factor is the presence of some sort of activator, while in the other it is enough to change the concentration of the solution.

On submergence (as a result of mountain building), the originally igneous and sedimentary rocks are altered at some critical depth and transformed into *metamorphic rocks*. Their formation is mainly due to the high temperature and pressure developed in the depths of the earth or in the vicinity of magmatic bodies. To this can be added the effect of hot underground waters and of gases emanating from the magma zone or from cooling magmatic rocks.

As for the distribution of the different types of rocks in the crust, calculations of thickness to a depth of 16-20 km give the following figures: 95% of igneous rocks, 4% of clay deposits, 0.75% of sandy formations and 0.25% of limestones.

CHAPTER II

A brief resume of the history of petrography

Petrography as a study of rocks, i.e., of mineral associations appearing as geologically independent bodies, naturally has its origins in both mineralogy and geology. The first concepts regarding rocks, as well as questions arising from their study, are revealed in old treatises on geology and mineralogy.

Petrography became a completely independent science with its own problems and methods of investigation at the beginning of the second half of the XIX century, as a result of the application of the methods of microscopy, and later of physical chemistry. It is only since then that it has passed from qualitative to quantitative determinations, i.e., it now operates with measurements and figures and, has become an exact science, making a real approach to the solution of its main problems—the development and constant alteration of rock types.

With the introduction of these methods, petrography is beginning to be seen as a strong influence, not only on the development of the other geological disciplines, especially those concerning mineral resources, but also on metallurgy, ceramics and glass production.

We shall now note some of the more important stages in the development of petrography.

By the end of the XVIIIth century, when mineralogy was already fairly well established, and was in possession of chemical, goniometric and partly crystal-optical methods of investigation, the development of petrography was still on a rather primitive level. It must, however, be pointed out that there had been some experimental work in petrography in the form of the melting of basalt and similar rocks in France and Great Britain at the end of the XVIIIth century and the very beginning of the XIXth century. Mitscherlich made mineralogical experiments a little later (in the twenties of the XIXth century); he was the first to describe certain slags and other industrial minerals. Subsequently, Senarmont

achieved quite considerable success in the systematic synthesis of various minerals. He heated solutions in sealed tubes and obtained artificial quartz, corundum, many sulphides and certain other minerals.

In the first decade of the second half of the XIXth century, a remarkable event took place in the history of rock science and metallography: the microscope began to be systematically applied to the study of rocks and metal alloys. There had even been earlier experiments of a similar type. Thus, Cordier, in 1815, had undertaken a microscopic investigation of rock powders in transmitted light; later, Nicol invented his renowned polarizing appliance, the Nicol prism, and carried out a microscopic investigation of sections of silicified wood in polarized light; in 1831, the Russian scientist P. Anosov succeeded in using the microscope in a study of a polished surface of cast and forged steel, after etching it with acids. The work of Sorby in England introduced a wider use of the microscope in the study of rocks: in 1858, by the microscopic analysis of inclusions in rock minerals, as an answer to petrogenetic questions, and in 1863, by an analysis of meteoritic and industrial iron.

In the XIXth century and at the beginning of the XXth century, the greatest attention of petrographers was concentrated on the study of magmatic rocks, and, in part, on questions of metamorphism, but there was an almost complete absence of any investigation of sedimentary formations. At the end of last century, petrographers had studied the igneous and metamorphic rocks in detail as to composition, but only as far as their main constituents, the silicates, were concerned. The remaining constituents, though of great significance in any understanding of ore formation had been studied only superficially. The cause of the scant attention paid to the sedimentary rocks was in part due to the extreme dispersion of their constituent particles, especially the clay particles, which could not be investigated at the small magnifications available in microscopes of that time.

The size and scope of this book do not permit a detailed exposition of the general history of petrography, and so only the main stages of its development in the U.S.S.R. will be given in what follows.

One of the first geological works written from the petrographical aspect to be published in Russia, was the work by M. Lomonosov, entitled *The Layers of the Earth* (1742). Soviet Academician V. Vernadskii called this work "the first brilliant outline of geological science in all the literature of the XVIIIth century, Russian and foreign". In this work, Lomonosov distinguishes secondary friable rocks (soil layers, sands and other detrital rocks of various grain size, clays, peat); secondary compact rocks (sandstones, limestones, shales); the primary rocks having their origin in the "heart of the earth".

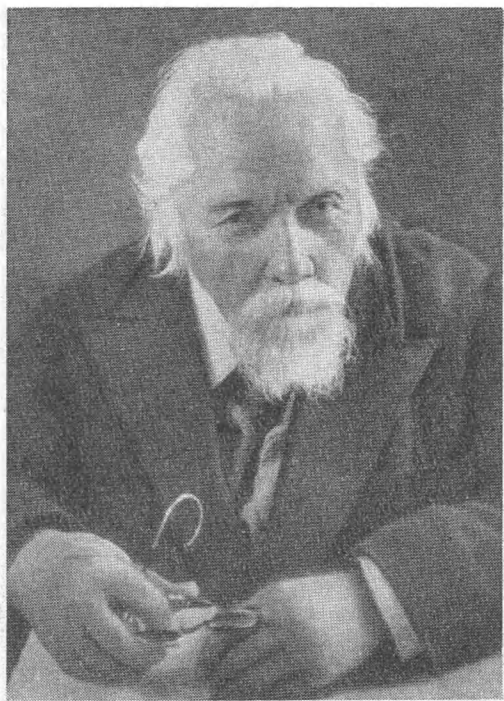


M. LOMONOSOV

At the end of the XVIIIth and the beginning of the XIXth century, the public lectures and many papers of Academician V. Severgin were of especial importance for mineralogy and petrography in general, and for the mineralogy and petrography of Russia in particular. He wrote the first Russian textbook of mineralogy, one of the sections of which "Folded Rocks" was specially devoted to petrography, being a classification of rocks.

The appearance of the petrological microscope changed the direction of petrographic work, and in 1867, A. Inostrantsev, and in 1868, Academician A. Karpinskii had produced thoroughly sound works on the microscopy of rocks.

As well as the detailed petrographic description of rocks, Karpinskii was also interested in questions concerning their formation and alteration (the order of crystallization of minerals from a magma,



A. KARPINSKII

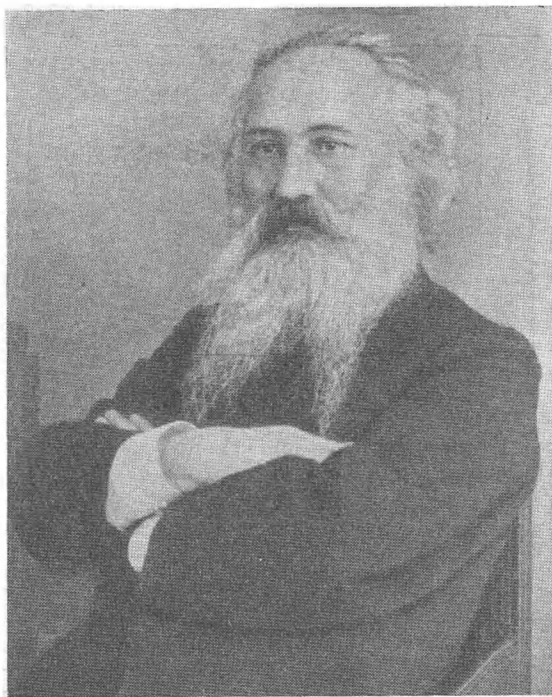
metamorphism). He was the first in Russia to compile a textbook of rocks.

The first fundamental researches of E. Fedorov and F. Levinson-Lessing appeared during the last two decades of the XIXth century, and a Russian petrographic school was founded.

Fedorov was the founder of present-day crystallography, and perfected the method of microscopic investigation; he was the creator of the theodolite method of investigating minerals by means of the Fedorov stage, and worked out the theory and practice of determining the composition of isomorphous mixtures of minerals from their optical properties. He was also interested in questions of petrographic nomenclature and rock chemistry.

Levinson-Lessing studied the causes of the variety in rock types, the solidification processes of the magma, the origin of granites, the intrusion mechanism of trap rocks, and made petrographic studies of a number of regions of Russia.

He was the first to apply experiment and make use of analogies with industrial processes in solving problems of magmatic petrog-



E. FEDOROV

raphy. The Kamchatka Vulcanological Station was created under his direction for the constant observation of active volcanoes.

The progress in physical chemistry (at the end of the XIXth and the beginning of the XXth century) created a foundation for deepening the knowledge of the laws governing the genesis and formation of rocks. The bias towards physical chemistry, along with the study of geological relationships, form the basis of present-day petrography.

An important aspect of the development of petrography in the Soviet Union is a renewal of experimental work based on the great advances made in physical chemistry and carried out under improved laboratory conditions. A number of petrographic scientific centres have appeared—the Levinson-Lessing Laboratory in the Polytechnical Institute in Leningrad, the experimental laboratory of the Academy of Sciences, the Leningrad Mining Institute, etc.

Much work is continuing to be published by Soviet petrographers. In particular, petrochemistry has been developed by A. Zavaritskii, technical petrography by D. Belyankin, the physical chem-



A. ZAVARITSKII

istry basis of metamorphism by D. Korzhinskii, petrography as the basis of contemporary metallogenesis by S. Smirnov and Yu. Bilibin, and a qualitative mineralogical classification of igneous rocks by V. Kupletskii. The work of a Moscow group of petrographers, E. Kusnetsov, V. Koptev-Dvornikov and others, is of great importance in the field of magmatic petrography.

The study of sedimentary rocks has followed a somewhat different path. It was conceived at almost the same time as magmatic petrography, but then received comparatively little attention, and the greatest development in sedimentary petrography has occurred only recently. Ya. Samoilov was one of the pioneers of this science. The works of V. Vernadskii (the role of living substance), A. Fersman (the geochemistry of sedimentary processes), and V. Baturin (the composition of oil-bearing beds) are of great importance.

The work of Academician A. Arkhangel'skii has been of tremendous importance in the development of the petrography of sedimentary rocks: he studied the wider aspects of the genetic relationships of chalk, bauxites and iron ores. Of especial interest are his comparative lithological studies of Black Sea sediments, of the origin of oil in the Northern Caucasus and of the formation of bauxite.

Academician N. Strakhov has continued Arkhangel'skii's work and developed his original ideas in numerous papers setting forth the historico-geological concepts controlling the processes of sedi-



D. BELYANKIN

mentation and developing comparative lithology as a method of investigation.

The petrography of sedimentary rocks is developing in many directions. Some investigations are made from the point of view of geochemistry or chemistry, others from that of paleontology and paleogeography, a third group is connected with the study of contemporary geological processes, a fourth with mineralogical analysis, a fifth with geology and stratigraphy, and a final group with the study of mineral resources; here, the main role is played by a study of the structure and texture of rocks. One particular trend is concerned with the problems of engineering geology and the study of mine waters.

F. Levinson-Lessing applied petrographic concepts to the solution of problems in engineering geology, and to the recognition of deposits of useful minerals, particularly structural materials. This work is now being continued by his pupil B. Zalesskii.

A special concern of petrography is the structural analysis of magmatic rocks. These problems are raised in the universally known work of Sander, Cloos, Balk, N. Yeliseyev and Pek in the U.S.S.R.

Advances in petrography are very closely connected with mining. Petrography cannot be separated from the study of deposits of useful minerals whether metallic or non-metallic, and the development of these sciences proceeds simultaneously. Each achievement in any one of them advances the whole complex of knowledge of rocks.

CHAPTER III

Some facts concerning the crystal chemistry of minerals

Most of the subject matter of this chapter should be known already to students from their chemistry, mineralogy and crystallography courses, and so it can be confined to a summary of those questions which apply especially to petrography.

One of the tasks of crystal chemistry is a detailed study of the internal structure of crystals and the elucidation of the relationship between these internal arrangements and chemical composition.

The charge on the nucleus is determined by the position of the element in the periodic table, i.e., by its atomic number. The number of electrons around the nucleus is equal to this number.

The elementary particles occupying the points in the space lattice of a crystal structure can be either atoms, ions or molecules. These elementary particles are termed *structural units*.

Polymorphism is the name given to the ability of one chemical substance to form two or more crystalline phases. Such substances are termed polymorphic modifications of the given compound or element. *

These will be individual minerals, each possessing its own peculiar properties, especially the optical ones.

If a substance forms a number of different polymorphic modifications, distinction is made between dimorphism, trimorphism and so on.

Any crystalline substance can be converted into any polymorphic modification by the appropriate variation of the thermodynamical conditions. The following rule then applies: the modification of a given substance corresponding to higher temperatures and pressures is usually of higher symmetry; thus, the high-temperature modifications of SiO_2 (Table 1), CaCO_3 , KClO_4 and other substances have a higher symmetry than the low-temperature forms.

* In the case of simple substances (elements), polymorphism is often referred to as allotropy.

Table 1

Modifications of SiO_2

Mineral	System	Specific gravity	N_m or N_o	Double refraction
Quartz	Trigonal	2.65	1.544	0.009
Tridymite	Orthorhombic	2.27	1.469	0.004
Cristobalite	Tetragonal (?)	2.35	1.487	0.003

Isomorphism. Isomorphism is the term given to the ability of compounds differing in chemical composition to give crystals in which the structural units of one substance can be replaced in varying proportions by structural units of the other substance. Such replacements are possible only in those cases when the crystals of the substances concerned have geometrically related structures, and the structural units themselves are similar in size, form and the type of chemical bonding.

As a consequence of the closeness in size of their structural units and of their chemical similarity, certain substances manifest an ability to form crystals of mixed composition: these are termed isomorphous mixtures.

The plagioclases, rock-forming minerals of wide distribution, are an interesting example of perfect isomorphism. They crystallize in various types of rocks, forming minerals of identical structure but different composition, and present a continuous series of mixed crystals of the two minerals—albite ($\text{NaAlSi}_3\text{O}_8$) and anorthite ($\text{CaAl}_2\text{Si}_2\text{O}_8$). All intermediate members of this series are found in rocks.

When the valencies of the members of an isomorphous pair are different, but the total valency is preserved, the isomorphism is termed *heterovalent*. For example, forsterite, Mg_2SiO_4 , and fayalite, Fe_2SiO_4 , mix in all proportions, forming homogeneous crystals.

When mixed crystals of limited miscibility are formed, strict observance of the structural similarity rule is not required. This type of isomorphism is thus often to be encountered in any study of crystals formed under natural conditions from systems containing widely different ingredients. Thus, orthoclase, KAlSi_3O_8 , and albite, $\text{NaAlSi}_3\text{O}_8$, give homogeneous mixed crystals only for certain small concentrations of the additive component.

The laws of isomorphism are essential both for an understanding of crystal structures and for the elucidation of the geochemical association of the elements in the different types of rocks. The study of isomorphism links crystal chemistry with geochemistry, petrography and mineralogy, in the most intimate way.

CHAPTER IV

Some crystallographic information

Rocks are made up of various minerals, which are the manifestations of crystalline substances, and the study of rocks consists mainly in the determination of these minerals and their mutual relationships. Every student of petrography must therefore have some knowledge of elementary crystallography. On the assumption that the reader already knows something of this subject, only certain definitions will now be given.

Crystalline substances are solid substances, in which the elementary particles are arranged in a definite pattern, forming a 'space lattice'. Usually they are polyhedra, the faces coinciding with the lattice planes, the edges with the densest rows of elementary particles, and the angles with the intersections of such rows. Solid substances in which the elementary particles are arranged at random are said to be amorphous. This structure determines the most important property of a crystalline substance, its *anisotropy*, which means that such a substance, having a uniform chemical composition, has different physical properties in different directions. This phenomenon is especially well manifested in the regular variation of the optical properties of a crystal. One of the manifestations of anisotropy is the anisotropy of growth, i.e., the ability of a substance to build crystals in the form of regular polyhedra of strictly defined form, inherent in the parameters of the space lattice.

A homogeneous substance whose properties are the same in all directions is said to be *isotropic*.

All crystalline substances possess symmetry, i.e., a regular arrangement of their particles in space.

This regularity is determined by the *symmetry elements*, which are fixed in a certain position for a given crystal, and determine not only its external form, but also its physical properties. The most important of the symmetry elements are:

(a) *centre of symmetry*, or as it is more often called, a centre of inversion (*C*); this is a conventional point inside the crystal. Any

straight line passing through it in either direction meets similar points of the figure at equal distances;

(b) *symmetry axis* (L) is a straight line, rotation around which repeats equivalent parts of the figure a whole number of times. This number gives the corresponding name to the symmetry axis. In crystals it is possible to have axes of twofold or higher symmetry, threefold, fourfold or sixfold;

(c) *symmetry plane* (P) divides the figure into two equal parts, which can be superposed by any simple means of mirror reflections.

There can be only one centre of inversion in a crystal, and different numbers of axes and planes, according to the requirements of the crystal form.

The complete assemblage of all the symmetry elements of a crystal is called the form or class of symmetry, and there are 32 such classes. The symmetry classes are combined into seven larger groups called *systems*.

When describing and representing the forms of crystals, especially in regard to the mutual dispositions of their faces in space, it is convenient to adopt, as is done in analytical geometry, certain directions passing through the centre of the ideal form, as coordinate axes. Their directions are determined by the crystal symmetry to a certain extent; in most cases symmetry axes are considered to be coordinate axes.

The lengths of the axes to their intersection by the faces are called the *parameters*. A certain face intersecting all three coordinate axes is selected as the *unit face*. The lengths of the axes to the intersection of this unit face on the coordinate axes serve as a scale for the corresponding axis. The crystal edges or the directions parallel to them, which have been adopted as the coordinate axes, are called the *crystallographic axes*. The angles between the crystallographic axes are designated: α —between the second and third axes; β —between the first and third; γ —between the first and second axes.

The ratios of the intersections of the unit plane, a , b , c , and the angles α , β and γ are called the *crystal elements* or the *geometric constants* of the crystal, and they are completely definitive of each system.

The position of each face in space will be determined by three integers expressing the ratios as fractions, the numerators of which are the parameters of the unit face, while the denominators are the parameters of the face with which we are concerned. These numbers are called the *symbols* of the face. The symbol of the unit face will be (111); the symbol of a face parallel to any crystallographic axis includes the index equal to zero corresponding to that axis, for example, (101) and so on.

The choice of crystallographic axes and unit face are known as the reference axes of the crystal (Fig. 1).

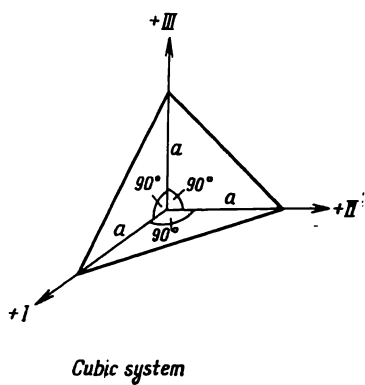
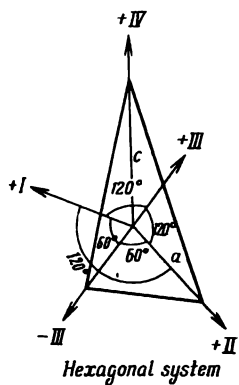
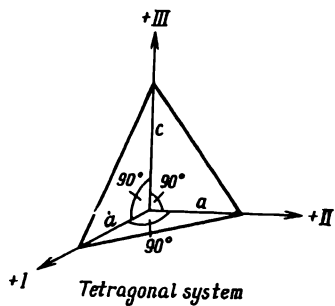
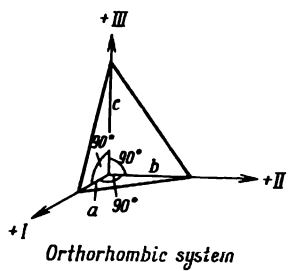
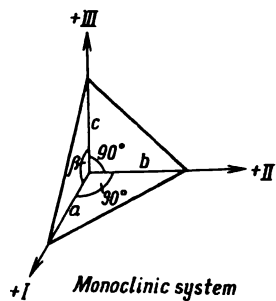
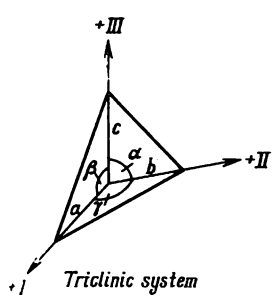


Fig. 1. Reference axes of the crystal systems

The following systems are distinguished:

Cubic—all crystals belonging to this system possess several axes of the highest order of symmetry, and must have four axes of threefold symmetry ($4L_3$) and three equal crystallographic axes of lengths $a : a : a$, arranged at right angles to each other (and coinciding with the axes of fourfold or twofold symmetry). The most important simple forms are the cube, octahedron and tetrahedron.

Tetragonal. In crystals of this system there is one axis of fourfold symmetry; three crystallographic axes at right angles to each other, two of which are equal in length (coinciding with twofold symmetry axes), and one of different length taken as the vertical axis (coinciding with the axis of fourfold symmetry) $a : a : c$. The most important simple forms are the prism, the pyramid and the dipyrmaid.

Trigonal and *hexagonal* are characterized by the presence of one axis of threefold symmetry (trigonal axis) or of sixfold symmetry (hexagonal). Three equal crystallographic axes are arranged in one plane at an angle of 60° (coinciding with the axes of twofold symmetry), and one vertical axis of different length perpendicular to the plane of the three equal axes and coincident with the axis of threefold or sixfold symmetry, $a : a : a : c$. The most important simple forms are the prism, pyramid, dipyrmaid, rhombohedron.

Orthorhombic has no axes of high symmetry. Three unequal crystallographic axes $a : b : c$ are at right angles to each other. If the c axis is taken as the vertical axis, then the longer lateral axis, b axis, will be directed to the right and to the left, and the shorter a axis from the front to the back. The most important simple forms are prisms and pinacoids.

Monoclinic. Three unequal crystallographic axes: one (coincides with the axis of twofold symmetry which is single in this system) is at right angles to the other two which are arranged at an oblique angle to each other. The axis normal to the other two is called the b axis. In order to orient the crystal the b axis is directed to the right and to the left while the c axis remains vertical. The most important simple forms are prisms and pinacoids.

Triclinic. This system is characterized by the absence of axes and plane of symmetry; three unequal crystallographic axes $a : b : c$ are arranged obliquely to each other at angles of α, β, γ . The main forms are pinacoids.

Crystals of any substance always reveal some variety in the relative development of their faces. This determines the *habit* (or, in other words, the aspect) of crystals. This causes no variation whatever in the angular relationships of the faces, nor, of course, in their symmetry.

If all faces are of approximately the same size, the crystal is termed *equidimensional* (and has isometric, or normal habit).

When two opposite faces are considerably larger than the other faces, the crystal appears to be flattened, and is said to be *tabular*.

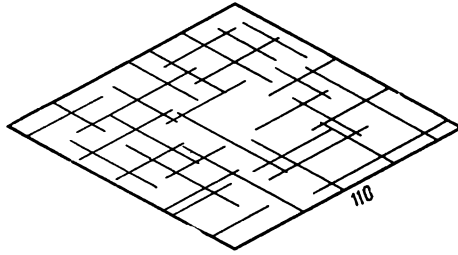


Fig. 2. Cleavage of amphibole in transverse section perpendicular to both cleavage planes

But when this flattening is considerable, the crystals are termed *platy* or *foliate*. If three or more faces belonging to one zone, i.e., parallel to any one line, are noticeably bigger than other faces, then the crystal will be elongated in the direction of this line and is termed *columnar*. The extension in a given direction can be so great that crystals become *acicular* or *fibrous*.

Some crystals show both lengthening and flattening: they are then termed *lathlike* or *rodlike*.

Cubic crystals are usually equidimensional; *tetragonal* and *hexagonal* ones are equidimensional, flattened normal to the vertical axis, or elongated parallel to it. Representatives of the other systems can be flattened parallel to any pinacoid or elongated parallel to any crystallographic axis.

Crystals often grow in groups (as druses). If they grow so that some analogous directions (axes) are precisely parallel, while others are not, then such intergrowths are called *twins*.

Crystals of isomorphous mixtures, whose composition changes from the centre to the periphery, are said to be *zoned*.

To determine the crystallographic attribution of grains without any visible traces of faces, the cleavage, twinning and optical properties are made use of.

Cleavage is the ability of a crystalline substance to split along strictly defined *planes*, corresponding to one or other of its faces. There can be a single cleavage direction (for example, along the pinacoid in mica) or several (for example, three along the cube faces in rock salt).

This phenomenon is explained by the fact that in the crystalline lattice of the mineral there exists a direction of greatest cohesion of ions and one of least cohesion, and it is along these relatively weak directions that the splitting of the mineral takes place.

Differently oriented sections through the one crystal reveal a different number of cleavage cracks, arranged differently in the different sections.

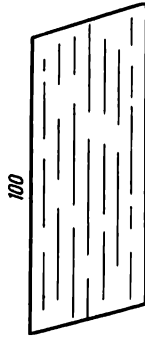


Fig. 3. Cleavage of amphibole in a section parallel to the prism

Thus, in an amphibole crystal, with prismatic cleavage in sections perpendicular to both cleavage planes (perpendicular to a prism), two well developed systems of narrow straight parallel cracks are seen, forming an angle of 124° between them (Fig. 2).

In sections parallel to a prism (Fig. 3), the cleavage cracks are parallel.

Under the microscope, cleavage is best seen in minerals with high refractivity of light. It is least well seen in sections normal to the cleavage planes.

Along with cleavage cracks which are usually discontinuous, cracks which pass uninterrupted across the whole grain can sometimes be seen in crystals. These cracks which do not always maintain a plane surface appear to be the result of external pressure, and are termed *partings*.

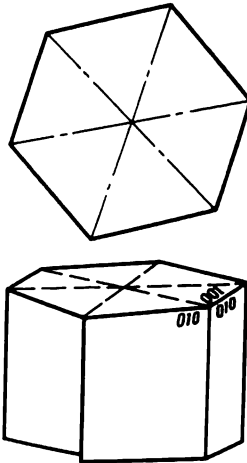


Fig. 4. Pseudohexagonal cyclic twin (trilling) of cordierite

Cracks which are irregular, uneven, twisting and without any conformity to the crystallographic directions of the crystal, comprise the *fracture* of the mineral. Minerals without any cleavage are characterized by conchoidal fracture.

Twinning is a regular intergrowth of crystals of one and the same crystalline substance, in which one individual can be derived from the other by simple geometric transformations such as rotation and reflection. As a result, twins possess a number of additional crystallographic elements, by means of which their symmetry can be determined. In the simplest case, the regular formation of twins is reflected in the fact that one of the two superimposed individuals might be turned from its original parallel position around a certain line, the so-called twinning axis, most often through 180° , after which it is intergrown with the second individual. The twinning axis is either a possible edge, or is perpendicular to a possible face, which is the twinning plane of the crystal. This is the plane, in which one individual of the twin can be made to coincide by reflection with the other or arranged in a parallel position if the intergrown crystals are unequal in size.

The plane along which the twinned crystals are united is called the *composition plane*, or the *twin suture*. It either coincides with the twin plane, or is parallel to the twin axis.

If twins are composed of several individuals, and the mutual relationships of each two adjacent individuals are the same, then this twinning is termed *polysynthetic*, when the composition planes are parallel, and *cyclic*, when these planes are turned full circle; in this case, there is often created a pseudosymmetry which is higher than the symmetry of the crystal, for example, orthorhombic cordierite appears to be hexagonal (Fig. 4).

Crystals of one and the same substance and structure can form intergrowths after different laws, i.e., they can have different twinning axes and planes, whereupon the composition planes will be different in each case.

Polysynthetic twinning with intergrowths according to more than one law is called complex twinning.

Twins formed artificially (during shearing or other stresses) are termed *secondary* (twins of calcite, complex twins of feldspars and of pyroxenes).

Each individual of a twin has its own crystallographic orientation and its own independent optics.

CHAPTER V

The elements of crystal optics

Every crystal of given composition and structure is characterized by definite optical constants by which the substance can be determined with sufficient accuracy without chemical analysis. Analysis by crystal optics, or, in other words, the optical method of determining a substance has now been well worked out and is widely used in mineralogy, petrography and chemistry.

1. ORDINARY AND POLARIZED LIGHT

Light is known to have a dual nature. On the one hand, it can present itself as a stream of light quanta or photons, and on the other, as undulating electromagnetic vibrations. In certain phenomena, the quantum nature of light is most obvious (luminescence, photo-effect), while in others light manifests its wave motion (refraction, reflection). When light is transmitted through a crystal, it is the undulatory nature of light that is most strongly manifested, which later will therefore be discussed in detail.

To bring to mind the process of harmonic electromagnetic vibration, let us take the projection of the uniform motion of a point around the circumference of a circle on to straight lines perpendicular to the ray passing through the centre of the circle (Fig. 5). We will thus obtain a combined representation of the motion of the point P simultaneously along the circle and along the straight line.

The angle α through which the point P has been displaced to position P_1 (Fig. 5), and α_1 to P_3 , is called the *phase* of the wave motion.

The time T taken by the point to pass along the entire length of the circumference, or the path $P-P$ along the straight line is called the *period* of the vibration. The inverse value $1/T$ is called the *vibration frequency* and determines the colour of the light.

The greatest distance of the point from the origin OP_2 and OP_6 is called the *amplitude* A of the vibration. The amplitude characterizes the intensity of light.

The distance through which the point passes along the line during one period of vibration is called the wavelength λ .

The velocity of propagation of a light wave in airless space $v = 300,000$ km/sec. It decreases when space is occupied by a substance and we then speak of an optically denser or an optically rarer medium. The wavelength $\lambda = vT$.

Let it be supposed that T is kept constant when passing from one medium to another. But if the ray of light which is being propagated in one medium meets the surface forming the boundary between it and the other medium, it will to some extent be partially reflected back into the first medium. The reflection of light from a plane surface obeys two laws: the angle of incidence is equal to the angle of reflection; the incident and reflected rays lie in one plane with the perpendicular raised from the point on the plane where the ray intersects it.

Another portion of the vibrations penetrate into the second medium, and as a consequence of the change in velocity of the propagation of light, the direction of the ray in the second medium alters: this phenomenon is called the refraction of light. It obeys two Descartes' laws: the incident and refracted rays lie in one plane with the perpendicular at the point of incidence of the ray to the plane on to which the ray is falling; the ratio of the sine of the incident angle to the sine of the refracted angle is a constant, which is called the refractive index n of the second medium in relation to the first (Fig. 6):

$$\frac{\sin i}{\sin r} = n$$

If the first medium is air, whose refractive index relative to a vacuum is almost precisely equal to unity (1.000294), then n is the absolute refractive index of the second medium.

The value n is equal to the ratio of the velocity of propagation of light in the two media concerned: $n = \frac{v_1}{v_2}$; this equation shows that the greater the velocity of propagation of light in the second

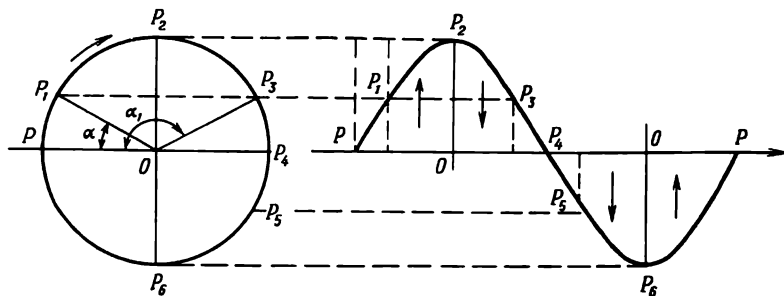


Fig. 5. Light vibrations

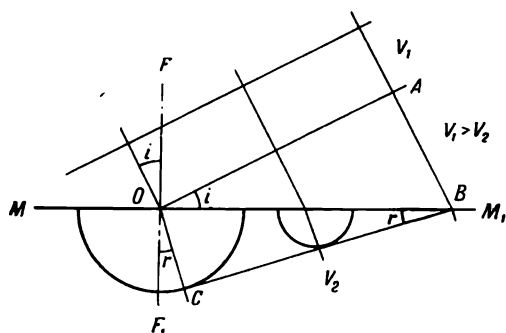


Fig. 6. Simple refraction

medium, the smaller is its refractive index. If $v_2 < v_1$, then n will be greater than unity; when $v_2 > v_1$, n is less than unity. There is thus a possible case, when the incident ray is wholly reflected from the plane of separation of the two media and will not pass into the other medium; this phenomenon bears the name of *total internal reflection*.

In a naturally occurring ray, the vibrations take place in all directions perpendicular to the direction of propagation of the light (Fig. 7a). In other words, the plane passing through the vibration and the direction of the ray is continually changing its position. As a result of reflection or refraction, the ordinary ray acquires a specific feature: the plane in which it is vibrating remains constant.

Such a ray is said to be *polarized*, or *plane polarized*. The plane, perpendicular to which the particles are vibrating, is called the *plane of polarization* (Fig. 7b). The reflected ray is polarized in the plane of incidence, the refracted ray in the plane perpendicular to this plane (Fig. 8).

In crystals belonging to the cubic system, and also in amorphous substances, light is propagated with the same velocity in all directions. For a ray of random direction in such a medium, all directions

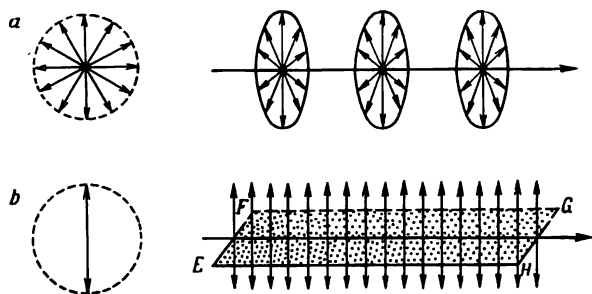


Fig. 7. Ordinary ray and polarized ray

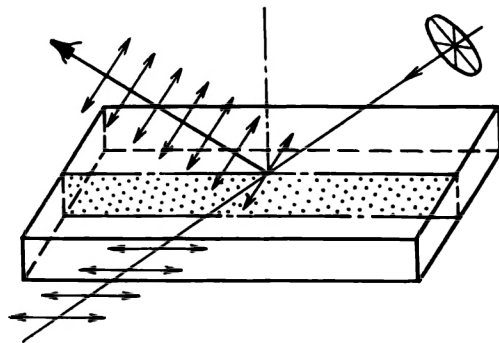


Fig. 8. Polarization by reflection and refraction

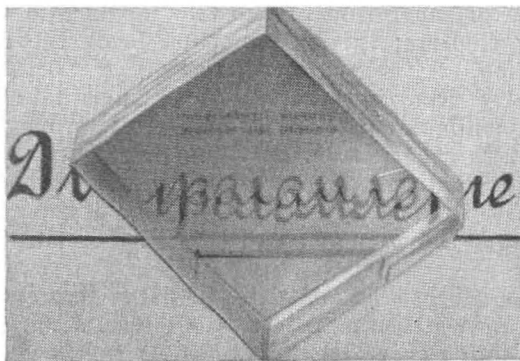


Fig. 9. Double refraction in Iceland spar

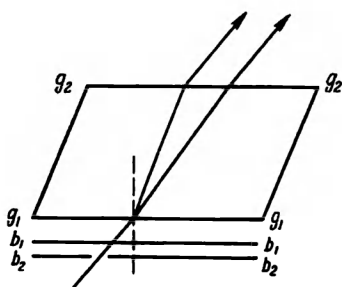


Fig. 10. Double refraction

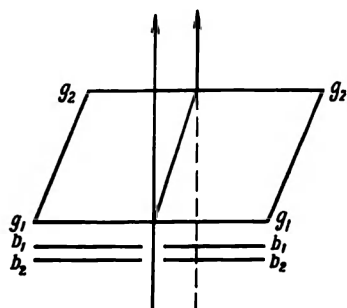


Fig. 11. Double refraction on perpendicular incidence of a beam of light

perpendicular to it are directions of possible vibration. In other words, cubic crystals and amorphous substances will have the same refractive index in any direction. Such substances are called *optically isotropic*.

In crystals belonging to other systems, the velocity of the propagation of light and, consequently, the refractive index and other optical properties will vary according to the section on to which the light is falling. Such substances are given the name of *optically anisotropic*. Thus, for example, if we take a piece of Iceland spar and look at an object (drawing) through it, this will appear to be double (Fig. 9). If a narrow pencil of rays falls in any arbitrary oblique direction on to a cleavage plane (Fig. 10), then it will be seen on the opposite plane that this pencil of rays has been divided into two parallel beams. This can only be explained by the fact that one of the rays of the beam gives rise to two after refraction in Iceland spar, and that these two are propagated in different directions at different velocities. Even if the beam of light falls on the cleavage plane perpendicularly, as shown in Fig. 11, two rays will emerge from the crystal. This phenomenon is called *double refraction* or *birefringence*.

Investigation of the rays emerging from anisotropic substance shows that they will be polarized in two mutually perpendicular planes.

One ray will then have the same velocity in all directions and the same refractive index (the ordinary ray), and the second will change its velocity and refractive index in accordance with the section of the crystal (the extraordinary ray).

2. THE WAVE SURFACE

In a vacuum, water, glass and other isotropic substances, and in crystals belonging to the cubic system, light is propagated with the same velocity (v_0) in all directions and the surface it reaches after a certain interval of time is the surface of a sphere. Thus, the wave surface in a cubic crystal is spherical.

Crystals of the tetragonal and hexagonal systems possess double refraction, and so the wave surfaces of such crystals are double (Fig. 12); the surface of a sphere for the ordinary wave o and the surface of an ellipsoid of rotation for the extraordinary wave e . If the velocity of propagation of light for the extraordinary wave v_e is greater than for the ordinary ray v_o , then the sphere will be inscribed inside the ellipsoid (Fig. 12a); if the velocity is greater for the ordinary ray than for the extraordinary, the sphere will prove to be described around the ellipsoid (Fig. 12b). The velocities can be equal ($v_o = v_e$) only along the axis of rotation of the ellipsoid, that is

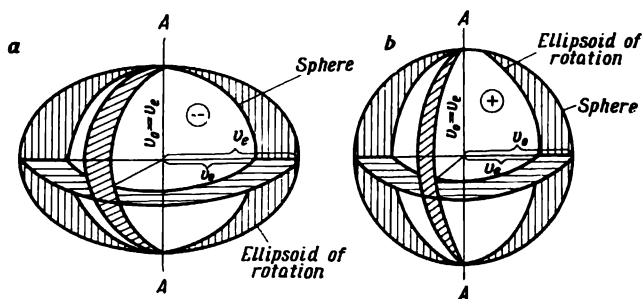


Fig. 12. Binary wave surface in crystals of the intermediate symmetry systems (tetragonal and hexagonal)

in the direction A which coincides with a unique direction in the crystal, i.e., with the axis of highest symmetry.

For orthorhombic, monoclinic and triclinic crystals, the wave surface is more complicated than in the case of cubic, hexagonal and tetragonal crystals. The velocity of propagation of both waves changes with the direction, and both waves are thus extraordinary.

In crystal optics it is customary to use not the double wave surfaces but single surfaces, each radius-vector of which shows the refractive index of the wave achieving its vibration in the direction of this radius-vector. This is what is known as the *optic indicatrix*. They are especially useful, because all the remaining optic properties of the crystal can be derived from them.

In cubic crystals the refractive index is the same in all directions, and so their indicatrices will have the form of a sphere. In hexagonal, trigonal and tetragonal crystals, there is one specific direction, in which lies the axis of highest symmetry, with the twofold symmetry axes perpendicular to it. The indicatrix of such crystals has the form of an ellipsoid of rotation, whose axis of rotation always coincides with the axis of highest symmetry. If the light passes through the crystal obliquely to the axis of rotation of the indicatrix, then in the section perpendicular to this direction there will be formed an ellipse, one axis of which will be equal to the diameter of the circular section of the ellipsoid, while the other to an intermediate value between the axis of rotation of the ellipsoid and the diameter of the circular section; this will then be the direction of vibration and refractive index of both the waves produced as the result of the double refraction.

The wave corresponding in refractive index to the circular section of the indicatrix and which is present in all its sections, will be the ordinary wave (designated n_o), while the other one whose value

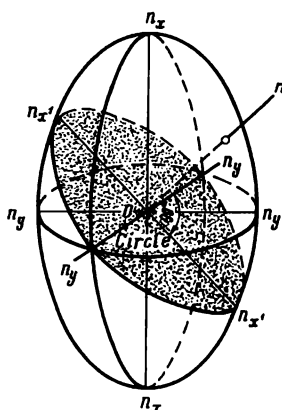


Fig. 13. Optical indicatrix of a uniaxial positive crystal

varies with the direction is the extraordinary one (n_e). If the ray of light is directed along the axis of rotation of the indicatrix, then the section of the indicatrix which is perpendicular to it will be a circle and consequently there will be no double refraction. Such a direction in a double refracting crystal, along which no double refraction is observed, is called an optic axis, and the plane passing through the axis of the optic indicatrix is called the principal section of the indicatrix.

In hexagonal, trigonal and tetragonal crystals, whose properties are represented by an ellipsoid of rotation, there is one optic axis, and they are therefore said to be *optically uniaxial*.

In some substances, the refractive index of the extraordinary ray is greater than that of the ordinary ray, in others, the converse is the case. The first substances are called *optically positive*, and the second ones *optically negative*.

The indicatrix of an optically positive crystal is an ellipsoid of rotation, extended along the optic axis (Fig. 13). For a uniaxial negative crystal, this ellipsoid of rotation is flattened along the optic axis (Fig. 14).

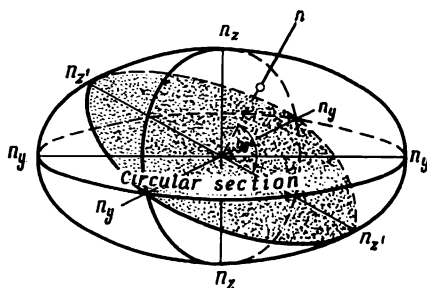


Fig. 14. Optical indicatrix of a uniaxial negative crystal

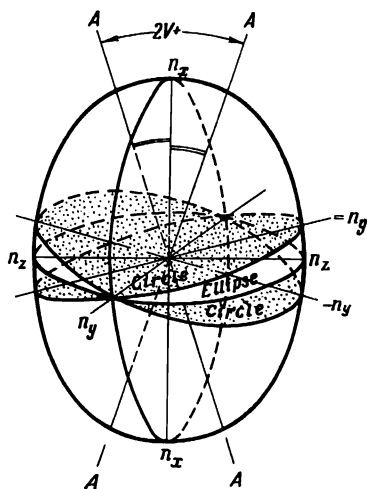


Fig. 15. Optical indicatrix of a biaxial positive crystal

In orthorhombic, monoclinic and triclinic crystals, the optical indicatrix will be a triaxial ellipsoid possessing three different axes corresponding to indices of refraction and designated n_x , n_y and n_z .

In virtue of the symmetry of a triaxial ellipsoid, two circular sections can be constructed through it with their diameters equal to its intermediate axis. Rays which are being propagated perpendicular to the circular sections, and which are consequently vibrating in the planes of these sections, will not undergo double refraction, and are thus the optic axes of the crystal. Since there are two such axes in these crystals, the crystals are said to be *optically*

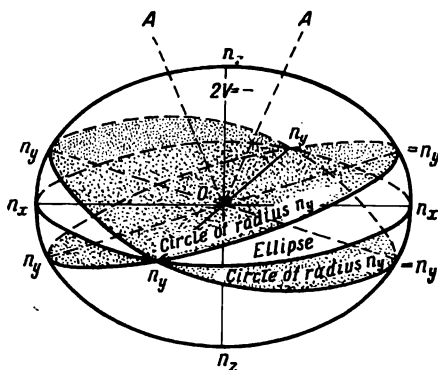


Fig. 16. Optical indicatrix of a biaxial negative crystal

biaxial. Both optic axes lie in that plane of the indicatrix in which occur its major and minor axes, and form two pairs of vertical angles.

The acute angle formed by the optic axes is called the *optic axial angle* and is designated by $2V$. The axis of the indicatrix bisecting the acute angle between the optic axes bears the name *acute bisectrix*, and that bisecting the obtuse angle the *obtuse bisectrix*.

The intermediate axis of the indicatrix perpendicular to the plane in which the optic axes lie is called the *optic normal*.

Two types of biaxial crystals are distinguished: one with its acute bisectrix coinciding with the greatest refractive index (n_x)—these are optically positive crystals (Fig. 15); in others, the direction of least refractive index (n_z) coincides with the acute bisectrix—these are optically negative crystals (Fig. 16). The optic normal is always n_y (the intermediate refractive index).

In orthorhombic crystals, their rectangular crystallographic axes coincide with the principal axes of the optical indicatrix n_x , n_y and n_z .

In monoclinic crystals, only one crystallographic axis, the b axis, is perpendicular to the two others, a and c , these latter intersecting each other at angles which are not right angles. Correspondingly, there is only a single axis of the indicatrix which coincides with the b axis and the two others do not coincide with either a or c .

In Fig. 17 is shown a section of a monoclinic crystal along its ac plane; the axis b is perpendicular to the drawing; the axis n_y of the indicatrix coincides with it; n_x and n_z lie in the ac plane.

In the triclinic system there is not a single right angle between the crystallographic axes; therefore, no axis of the indicatrix coincides with any crystallographic axis, but forms some angle more or less constant for each substance.

Biaxial crystals possess three principal sections: n_x-n_z , n_x-n_y and n_y-n_z .

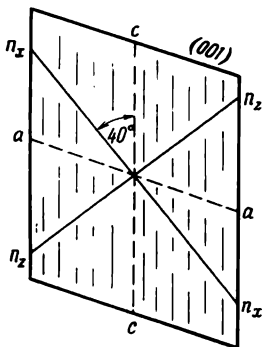


Fig. 17. The position of the optic indicatrix in monoclinic crystals

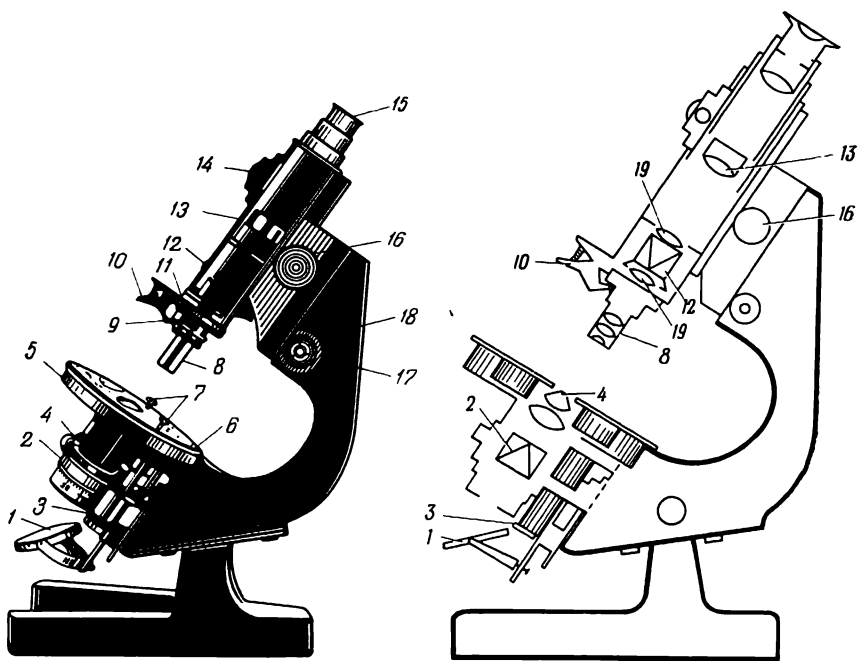


Fig. 18. External view of a laboratory model of a large polarizing microscope:

1—illuminating mirror; 2—condenser with a polarizing prism; 3—lever for moving out the condenser; 4—lever for insertion of the Lasaulx lens; 5—rotating stage; 6—vernier; 7—grips holding the section; 8—objective; 9—left centring screw; 10—clutches for attaching objective; 11—slit for compensator; 12—analyzer; 13—Bertrand lens; 14—screw for focusing Bertrand lens; 15—ocular; 16—attachment for rough focusing; 17—screw for micrometer movement; 18—vernier for micrometer movement; 19—correcting lenses

The polarizing microscope. In order to study the optical properties of anisotropic crystals it is necessary to use light with a definite vibration direction, i.e., plane-polarized light. A special microscope is used for this and can at the same time be used to see the structural details of minerals or rocks. The most important part of such a microscope are two similar fittings, one of which, the lower, giving polarized light is called the polarizer, and the other, the upper sliding one, which can be used to study the light obtained after passage through a crystal plate, and which is called the analyzer.

These devices can be either special polarizing prisms made out of Iceland spar, in which one of the two rays obtained as a result of double refraction is removed from the field of view, or what is known as *polaroid*, which is usually a plate of celluloid in which have crystallized parallel crystals of a substance which has a very strong absorption for one of the two rays produced in it. The old

name of "Nicols" is retained regardless of the type of polarizer and analyzer actually used.

Besides the Nicols, such a microscope is fitted with a number of additional attachments, the principal ones being: a rotating stage, lighting apparatus, auxiliary lenses, oculars, objectives, various compensators, apparatus for the study of preparations in reflected light and the Fedorov universal stage.

The rotating stage is divided into degrees, has a vernier for the precise determination of rotation, is fitted with grips for the direct attachment of the preparation, and apertures to which can be attached the Fedorov stage or other supplementary apparatus. The stage rotates around the axis of the microscope and has a circular opening for the passage of light.

The ocular of a polarizing microscope is fitted with a cross made from cross hairs, one hair of which is arranged parallel to the direction of vibration in the polarizer, the other perpendicular to it; sometimes a micrometer is available for insertion or superposition.

The most up-to-date models of the polarizing microscope are of the highest optical standard and essential for the precise investigation of crystal optics. For teaching purposes, any widely distributed model is quite satisfactory, for example, that represented in Fig. 18.

With the polarizing microscope, the preparation can not only be examined under magnification, but it is a thoroughly precise measuring instrument, by means of which the physical constants of any substance can be determined, and hence the substance itself identified.

CHAPTER VI

Studying crystals by polarizing microscope

A rock which is to be studied under the microscope has a plate 0.02 to 0.04 mm thick prepared from it, and this is known as a thin section (Fig. 19). For this, a piece of the rock is polished on one side, and cemented by this side with melted Canada balsam*, *c*, to the object glass *d*. The opposite side is then polished. When the flat plate of rock is thin enough, the well boiled balsam is poured on to it and it is covered with a cover glass *a*.

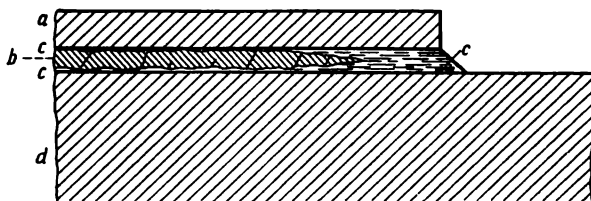


Fig. 19. Transverse section of slice (thin section)

Such a section can be studied in transmitted light. When the rock or mineral is being investigated in reflected light, only one side of the chip is cut and polished, and the entire chip is then fastened with a cement on to a cover glass. A one-sided section is known as a polished section and is mainly used to study opaque ore minerals.

1. THE DETERMINATION OF SIZE AND QUANTITY OF MINERAL GRAINS IN SECTION

A micrometer sited in the ocular is used to determine the size of mineral grains in section. The micrometer is fitted with a scale divided into 100 parts.

* The resin of the balsam fir, free of volatile oils, in its external form reminiscent of colophony (rosin). A similar substance obtained from the Siberian fir is used in the U.S.S.R., and is sometimes distinguished from Canada balsam by the name "fir balsam".

The ocular micrometer is calibrated beforehand, i.e., its divisions are evaluated against different objectives. A micrometer object is used for this, which is a glass plate having the form of a section with an inscribed scale 2 mm long, divided in its centre into 200 parts, each division being 0.01 mm.

The micrometer object is placed on the microscope stage in such a way that its zero point coincides with the zero point of the scale in the ocular, and the number of divisions of the ocular corresponding to 2 mm are noted. The value of one division of the ocular can be determined by simple division into 100, after which measurement of the size of a grain presents no further difficulty.

The quantitative mineralogical composition of a section is determined by using an ocular with a square network micrometer; this can be used because the quantitative ratios of the areas, occupied in the section by all grains of each individual mineral, are equal to the ratios of these minerals in the rock.

Special devices are employed for more precise calculations: they allow a mechanical summation of the area of the section occupied by any mineral.

2. STUDY OF SECTIONS WITHOUT ANALYZER

Before studying a section, it is first necessary to centre the microscope, i.e., to ensure that the axis of the microscope, as shown by the intersection of the cross hairs in the ocular, coincides with the axis of the objective and the axis of rotation of the stage. Any distinct mineral grain is placed at the intersection of the cross hairs and the stage rotated, whereupon the grain will describe a certain circle. Return the grain to the cross hairs, and turn the stage through 180°, after which by means of special centring screws situated on the objective the grain is moved through half the distance between it and the cross hairs intersection. By repeating this operation, a situation is reached when the grain stops moving away from the intersection. When the microscope is centred, the study of the section can be begun.

The form of the mineral. A rock is usually an aggregate of minerals which are seen in section in the polished slices. Minerals which have grown in their own crystal forms, with well developed traces of faces, are said to be *idiomorphic*; minerals only partially faceted, are said to be *hypidiomorphic*; minerals completely free of their characteristic faces are said to be *allotriomorphic*.

Idiomorphic grains have different outlines depending on their position in the slice; for example, prismatic hexagonal or tetragonal crystals in sections perpendicular to their major axis, have the form of a hexagon or square, while in sections parallel to the major axis, they are rectangular.

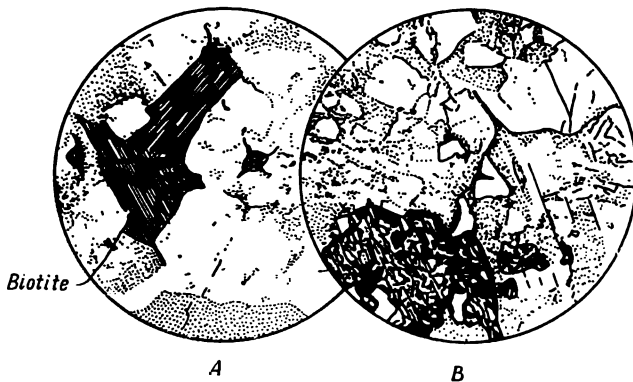


Fig. 20. The appearance of cleavage in thin section; $\times 20$:

A—biotite granite; two crystals of biotite with well-developed cleavage, feldspar and quartz can be seen; *B*—hornblende granite; brownish-green hornblende showing distinct cleavage in one direction, magnetite, oligoclase, orthoclase and quartz

Cleavage is especially clearly seen when the condenser is lowered or the diaphragm closed. In slice, the cleavage cracks are seen as a series of parallel straight lines. Thus, for example, mica has one system of cleavage cracks (Fig. 20), but sections of mica can be seen in slice showing no cleavage; amphiboles possess two cleavage directions with angles of $124^{\circ}30'$ and $55^{\circ}30'$ (see Fig. 2), augite—two directions of cleavage at an angle of 88° (almost a right angle) (Fig. 21).

Cleavage angles can be measured; to do this, the apex of the angle is placed at the centre of the field of view and the stage is rotated until one of the arms of the angle coincides with a cross hair in the ocular, and a first reading is made from the vernier of the rotating stage. The stage is then rotated until the other arm of the angle coincides with the same cross hair and a second reading is made. The measurement will be correct only when the section is perpendicular to the cleavage plane (in this case, when the microscope tube

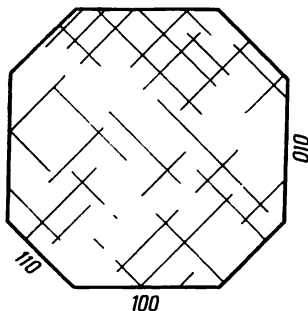


Fig. 21. Pyroxene cleavage

is raised or lowered, the cleavage cracks should stay in the same place without moving, and both systems of cleavage cracks should appear as fine straight lines). In any other section, the angle between the cleavage traces will be distorted.

The refractive index of minerals. A method of approximate determination of the refractive index of minerals in a thin section is based on a study of their type of surface. In a thin section, all minerals are covered above and below by a layer of Canada balsam with a refractive index of 1.54. At the same time, the surface of the section is always covered by fine furrows as a result of sectioning. If the refractive indices of the Canada balsam and the mineral are different, then the rays will be diverted at the boundary between them as a result of refraction and total internal reflection. These phenomena express themselves the more intensely, the greater the difference between the refractive indices of the balsam and the mineral, and the more the rays passing through the preparation are parallel to each other. Parallel rays are obtained by lowering the polarizer with the illuminating lens lying on it, or by narrowing the diaphragm. The preparation becomes darker, but the manifestation of refraction and reflection stand out more sharply. As a consequence of this diversion, not all the rays are seen by the eyes (through the microscope), and if the index of refraction is different from that of Canada balsam, the surface of the mineral seems to be covered with points merging into little wrinkles giving a shagreen-like surface.

The phenomenon of total internal reflection is often observed in just the same way at the boundary between two minerals with dissimilar refractive indices; as a result, the mineral with the greater refractive index is surrounded by a border which is the wider and sharper, the greater is the difference between its refractive index and that of its neighbouring mineral. It seems as if this mineral is standing out above the surrounding minerals, it possesses *strong* (Figs. 22 and 23) or *weak* (Fig. 24) *relief*.

The following method is used to determine the difference in refractive index of two adjacent minerals, or of a mineral and balsam at the edge of a slice. At the face between the two minerals (or the two media) even if the difference of refractive index is very slight, a fine band of light (the Becke line) will be seen. When the position of the microscope tube is altered, this line moves; the line movement indicates the relative magnitude of the light refraction of the contiguous media. Thus, when the tube is raised, the Becke line moves towards the mineral (or medium) with the greater refractive index. The appearance of this line is explained by the fact that part of the rays fall at a slanting angle on to the boundary between the two minerals, and, when they meet the mineral with the lesser refractive index, undergo total internal reflection.

The Becke line is seen distinctly, when the illumination cone is narrowed to an angle of total reflection. This is achieved by cutting

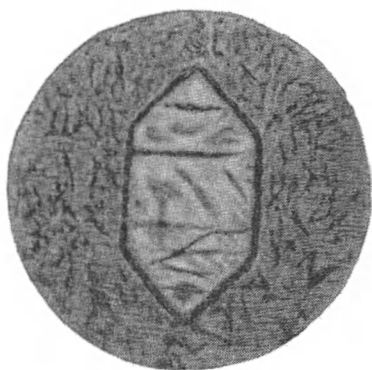


Fig. 22. Very strong relief
(garnet)

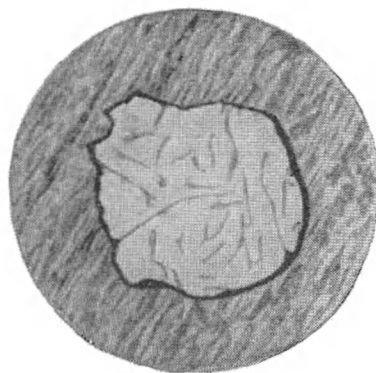


Fig. 23. Sharp relief (olivine)

off the appropriate rays by an iris diaphragm or by lowering the lighting system.

V. Lodochnikov recommends to study the refractive index of colourless minerals by what is known as the *dispersion effect*. Of two adjoining completely colourless minerals, the one with the higher refraction will appear a barely perceptible greenish colour in section, while the lower refracting will appear golden-yellowish. In order to accustom the eyes to the dispersion effect, V. Lodochnikov recommends that when working at the microscope, one should take a section of a fresh microcline granite and observe the colours in transmitted light of the microcline and quartz at their junction (this necessitates a slight closing of the diaphragm under the microscope stage, and a magnification of not less than $\times 20$). Once this effect has been noted for these minerals, it is easily seen at the boundaries of other transparent minerals.

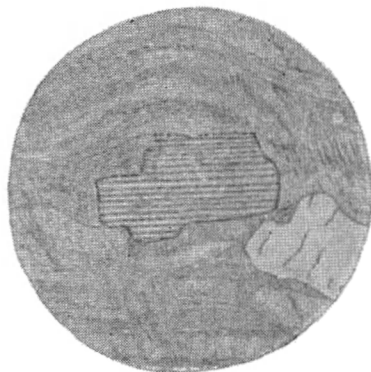


Fig. 24. Weak relief (biotite)

The possibility of judging the relative value of the light refraction of two adjoining media by means of the Becke line or any other method also makes it possible to determine accurately the light refraction of any mineral grain, even very small ones.

The immersion method has been developed for this purpose, which consists in submerging the grain under examination in a liquid whose refractive index is already known.

After this, the Becke line is used to compare the refraction of the mineral and the liquid. By subsequent selection of liquids, the refractive index of the mineral can ultimately be found to be intermediate between two adjacent liquids of a series, and will then be equal to the arithmetical mean of the refractive indices of these liquids. For the immersion method, sets of 100 liquids are made up with refractive indices n from 1.4 to 2.06, the refractive indices of neighbouring liquids differing from one another by 3-7 units in the third decimal place.

These sets are made up of mixtures of some of the following liquids:

	n		n
Water	1.336	Bromoform	1.590
Ethyl alcohol	1.360	Cinnamon oil	1.605
Glycerine	1.467	Monoiodobenzene	1.619
Castor oil	1.478	α -monochloronaphthalene	1.635
Xylene	1.487	α -monobromonaphthalene	1.655
Cedar oil	1.507	Methylene iodide	1.740
Monochlorbenzene . .	1.523	Sulphur in methylene iodide solu-	
Cloves oil	1.544	tion	1.800
Nitrobenzene	1.552	Solution of sulphur and phosphorus	
Monobromobenzene . .	1.561	in methylene iodide	2.06

Any sort of refractometer can be used to determine the light refraction of the liquids or for their calibration. In particular, the crystal refractometer represented in Fig. 25 is very useful. It is based on the principle of total internal reflection and can be used to determine the refraction of liquids as well as large crystals having one polished face.

Pleochroism. The colour of a mineral plays an important part in its determination. In sections of cubic crystals, the colour is constant, but the colour of doubly refracting minerals changes with the change of direction along which the light passing through the crystal is vibrating. The ability of minerals to absorb different amounts of light of different wave lengths when they are vibrating along different directions of the optic indicatrix is called *pleochroism*. It is obvious that there can be no pleochroism in crystals belonging to the cubic system, since their optic properties do not vary with direction. In crystals of the other systems, the absorption and tint change with the change of direction according to the

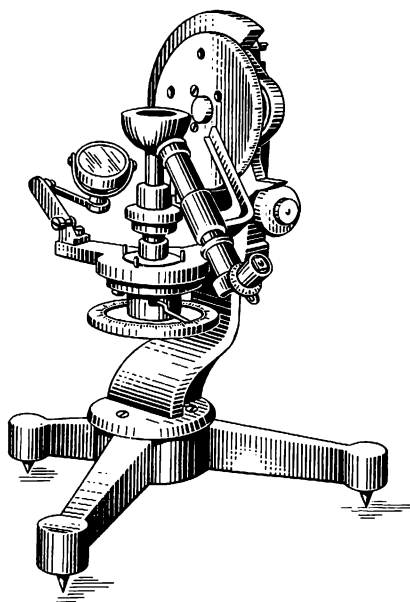


Fig. 25. Crystal refractometer

geometry of the ellipsoid of rotation for uniaxial crystals, and of the triaxial ellipsoid for biaxial crystals.

Pleochroism is observed under the microscope with the polarizer alone; when the microscope stage is rotated, first one and then another axis of the indicatrix is brought into coincidence with the vibration direction of the polarizer, and the tint obtained observed. Three types of phenomena are distinguished:

1. Various parts of the spectrum are absorbed in different directions and the intensity of the colour does not change (alkaline pyroxenes give light green and light yellow tints). This is *properly pleochroism*.

2. One and the same portion of the spectrum is absorbed in all directions, but with different intensity; thus, biotite, when the vibration in the polarizer is parallel to the cleavage, is coloured dark brown, but when the vibrations are perpendicular to the cleavage, pale brown. Here there is no change of colour (pleochroism), but a non-uniform absorption of light takes place, or what is known as *absorption*.

3. Different parts of the spectrum are absorbed with different intensity in different directions; both the tint and depth of colour of the mineral change (ordinary hornblende gives in some directions a dark green colour, in others, a light yellow). This is pleochroism and non-uniform absorption occurring together.

Pure colours corresponding to the axes of the ellipsoid are obtained when any one of them coincides with the vibration direction of light passing through the polarizer. In intermediate positions, the ray emerging from the polarizer dissociates in the crystal into two rays with vibrations along the axes of the ellipse. Since both these rays affect the eyes of the observer simultaneously, they give the impression of a compound intermediate tint.

Minerals belonging to the hexagonal and tetragonal systems can have only two clearly distinguishable absorption tints: one for the vibrations parallel to the long axis (the extraordinary ray), the other for vibrations perpendicular to it (the ordinary ray). The pleochroism formula for tourmaline, for example, is: n_o —light yellow, n_e —dark brown; absorption formula $n_e > n_o$.

In biaxial crystals there are three directions with different tints, and the pleochroism formula consists of three terms: for hornblende, for example, n_y —bluish-green, n_x —dark green, n_p —pale yellowish-green; absorption formula $n_x > n_y > n_z$.

The appearance of pleochroic haloes on rotation of the section with polarizer alone is closely connected with pleochroism. This phenomenon consists in that around inclusions in certain non-pleochroic or weakly pleochroic minerals there appear and disappear coloured circles, their greatest intensity of colour coinciding with the greatest intensity of colour of the mineral itself. They are most frequently observed around inclusions of zircon, orthite or other radioactive minerals; this confirms the conclusion that the cause of the appearance of these haloes is radioactivity.

3. INVESTIGATION BETWEEN CROSSED NICOLS IN PARALLEL LIGHT

The field of view will become dark if the vibration direction in the analyzer is arranged so that it is perpendicular to the vibration direction in the polarizer, and the analyzer is inserted into the tube of the microscope (crossed nicols). This is caused by the fact that the polarizer transmits only those rays which are vibrating in its principal section. The plane of the principal section of the analyzer is arranged perpendicular to the plane of the principal section of the polarizer, and this means that the ray which falls on to the analyzer is vibrating in a direction perpendicular to its principal section, and will therefore be absorbed.

If a plate of an isotropic mineral is placed on the microscope stage, the darkness of the field of view is not disturbed, because the isotropic mineral transmits rays with vibrations in all directions, and so the ray received from the polarizer passes through it without changing its vibration direction. It is otherwise when instead of the isotropic plate, an anisotropic plate is placed on the stage;

cancel each other out; in the intermediate situations, when two waves are out of phase, there is a certain resultant wave intermediate between its originators. There is accordingly a sharp intensification of light, a complete darkening or the formation of an intermediate degree of light.

If then a mineral plate placed between crossed nicols has the axes of its indicatrix parallel to the vibrations of the nicols, then the ray which has emerged from the polarizer will enter the crystal plate without any change in the direction of its vibration and, on emergence from the plate, will be absorbed by the analyzer. This will cause the darkening of the plate, or, as it is called, extinction. When the crystal plate is rotated on the microscope stage, this extinction takes place four times.

Anisotropic crystals may be easily distinguished from isotropic by this lighting up of the field of view and the presence of interference colours.

The determination of the thickness of a crystal and of the strength of the double refraction. The path difference which determines the interference tint of a crystal depends on the strength of the double refraction and the thickness of the slice. The simplest way to make a rough determination of the thickness is as follows. The microscope is adjusted so that the lower surface of the slice is clearly visible; the division of the micrometer screw against the static scale on the stand is noted; the micrometer screw is turned until the upper surface of the slice is clearly seen, and a second reading is made. From the thickness of the slice and the interference tint the strength of the double refraction can be determined, this being the most important constant of every anisotropic mineral.

This task can be carried out more rapidly by the *comparison method*. Coloured bands corresponding to the different interference tints of minerals are drawn vertically on a special chart. The colours are divided into several orders, the boundaries between which are determined by the position of the violet band. Horizontal bands show the thickness of the slice, while a bundle of lines diverging from the lower left corner of the table, and the figures on the top and to the right give the strength of the double refraction (Fig. 27).

Knowing the thickness of the slice and the tint, the point of intersection of the appropriate horizontal and vertical lines can be found. One of the radial diverging lines passes through this point, and the value of the double refraction is shown at the upper end of this line.

The thickness of the slice can be determined by using this table when the double refraction of the mineral is known. A grain of quartz is very often used for this, its double refraction being very constant and equal to 0.009.

If we are dealing with sections of a crystal which has very weak double refraction, it is difficult to decide whether such a section

has any effect on polarized light, i.e., whether the mineral is anisotropic (grey interference colours of the lowest order are difficult to distinguish from full darkness). In this case, the method of compensation of path difference is used to determine the double refraction. A special compensator is employed, a gypsum plate, for example, which gives the sensitive violet tint between crossed nicols. This plate is inserted into a slit in the microscope over the objective when the nicols are crossed. The faintest path difference in the section being studied is sufficient to change the colour of this plate and so to decide whether the crystal is doubly refracting.

It is useful to remember the double refraction and interference tint of the following, most common rock forming minerals (in sections of normal thickness).

Very weak (steel grey): leucite, apatite, nepheline.

Weak (bluish-grey to white of the first order): microcline, orthoclase, albite, oligoclase, andesine, labradorite, quartz, enstatite.

Medium (white, yellow or orange of the first order): andalusite, chlorite, anorthite, hypersthene.

Strong (red of the first order to violet and blue of the second order): tourmaline, augite, common hornblende and actinolite.

Very strong (green, yellow or orange of the second order): olivine, epidote, talc, biotite, muscovite.

Extremely strong (pale colours of the third and fourth orders to almost pure white): zircon, iron-rich hornblende, sphene, calcite and dolomite, rutile.

It should be noted that the interference colours of minerals such as the deep coloured micas and hornblendes which have strong absorption are masked by the actual tint of the mineral.

Extinction angles. Sections of anisotropic minerals, when viewed between crossed nicols, reveal an illumination or an interference tint which disappears four times during rotation of the microscope stage: the grain becomes dark or is extinguished. This takes place when the vibration direction transmitted by the crystal plate coincides with the vibration direction transmitted by the polarizer or analyzer. But vibration can take place in the crystal plate only in the direction of the major or minor axis of the indicatrix in the given section. By observing the extinction of the plate, the position of the axes of the indicatrix (n_x , n_y or n_z) can be determined as well as the angle formed by them with the traces of the crystal axes in this section of the crystal.

Straight or parallel extinction is an extinction which occurs when traces of the crystal axes of the given section of the crystal are parallel to the principal sections of the nicols. The directions of the crystal axes in the section are decided by its shape or by the position of cleavages in it (Fig. 28).

It is obvious that straight extinction is a sign that the crystal axes coincide with the axes of the indicatrix in that section of the

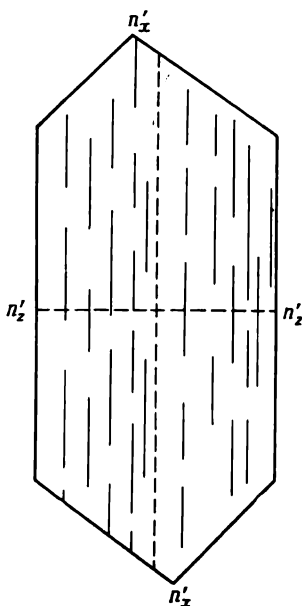


Fig. 28. Straight extinction in a longitudinal section of a monoclinic mineral (pyroxene or amphibole) along (100). The extinction is parallel to the traces of the vertical faces and to the perfect cleavage cracks along (110)

crystal. There must be straight extinction in uniaxial crystals and in any sections of orthorhombic crystals as well as in sections passing through the b axis of monoclinic crystals.

Oblique extinction (see Fig. 17) is seen when traces of the crystal axes in the given section do not coincide with the axes of the indicatrix, but are rotated through some angle relative to the principal sections of the nicols, and this angle is termed the extinction angle. Oblique extinction is characteristic of any sections of triclinic crystals, and for all those sections of monoclinic crystals which are not parallel to the b axis.

The size of the extinction angle in any one crystal form depends on the direction of the section. In monoclinic crystals it reaches its maximum value in sections perpendicular to the b axis (see Fig. 17, angle $cn_x = 40^\circ$).

Extinction angles vary with the composition of the crystalline substance, and the latter can be determined by its extinction angles.

The extinction angle is usually measured against the n_x axis (see Fig. 17). If, however, the extinction angles in different sections of a mineral are measured with reference to cleavage cracks, elongation, twinning, then the extinction angle of that section is spoken of as being with one of these elements.

To determine the extinction angle, use the analyzer, and look

for a grain of the mineral in question, which has a definite orientation, for example, a grain with the highest interference tint (the axes n_x and n_z lie in this section). This is brought to the intersection of the cross hairs in the ocular, then the microscope stage is rotated until some crystallographic direction (cleavage, or maybe a face) is brought into coincidence with one of the cross hairs. When the analyzer is inserted, there are two possibilities: the grain is extinguished (see Fig. 28), or the grain is not extinguished (see Fig. 17). The first is a case of straight extinction, the second of oblique extinction. In the second case, a reading is made of the position of the stage, and the preparation is rotated until it is extinguished in the direction of the least angle necessary, and another reading is taken: the difference in the readings gives the value of the extinction angle formed by the cleavage direction with one of the axes of the ellipsoid. There remains to be determined with which of these axes (n_x or n_z) the angle has been measured. How to determine this will be described later.

The extinction angle must be determined in several sections if it is to be used to determine a mineral. Thus, for example, sections of a hornblende crystal along various directions parallel to the vertical axis (there are two cleavages in these sections, but they give only one system of parallel lines) will give extinction angles varying from straight extinction in sections parallel to the pinacoid (100) to a maximum in a certain other section. This maximum angle is characteristic for a given mineral.

Optical character of the elongation. The minerals present in rocks are often elongated (or tabular) in certain crystallographic directions which coincide with the direction of greatest refraction of light n_x in some crystals, with the least refraction n_z in others. In the first case we speak of the positive elongation of the mineral, and the positive optical character of the *principal zone* (the name conventionally applied to the length of a mineral); in the second case we speak of negative elongation, and the negative character of the principal zone. The optical character of the principal zone or of the elongation is determined by making use of either a gypsum plate, a quartz wedge or other compensators, i.e., supplementary crystal plates, the orientation of whose indicatrices is already known. Their application is based on the fact that, if another double refracting mineral is placed above the section of the first one, and the second plate is oriented in the same way as the first, then the interference tint (the path difference) will be raised, as if being summed up; but if the direction of the greatest refraction of light n_x of one crystal coincides with the direction of the least n_z of the other, then the path difference is the result of subtraction and the tint is lowered.

The gypsum plate. A plate cut out of a crystal of gypsum is widely used as a compensator. Such a plate is now often made by polishing

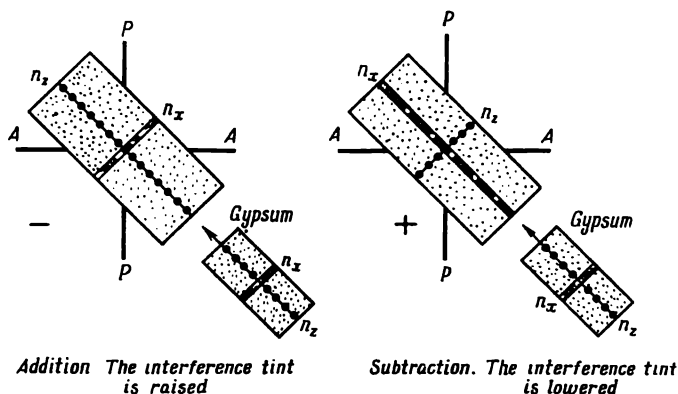


Fig. 29. Determination of the axes of the indicatrix and of the optical elongation using a gypsum plate

a suitably oriented quartz crystal, but the original name of the plate has been preserved. It is made of such a thickness that between crossed nicols it gives the sensitive violet or the violet red tint of the first order. If such a plate is introduced into the microscope, then the slightest change in the path difference of the rays passing through its system (for example, if a weakly doubly refracting crystal is placed on the stage) will cause either a yellow tint to appear in the ocular when the path difference is lowered, which occurs when the similar axes of the indicatrix of the gypsum plate and of the plate of the specimen mineral are crossed, or a blue tint if the path difference is increased, which occurs for a parallel arrangement of the same axes of the indicatrix. The direction of greatest refraction n_x is marked on the housing of the gypsum plate and usually coincides with the short side of the plate.

The determination of the character of elongation using the gypsum plate is carried out as follows: the mineral is first placed at extinction, and then the microscope stage is turned through 45° to obtain the greatest illumination (whereupon the axes of the indicatrix will take up a position diagonally with the nicols). After this, the gypsum plate is inserted into a slot in the microscope tube, and observations are made of the change of tint. If a lowering of the interference tint is observed, the axes of the same name in the indicatrix and gypsum plate do not coincide. If the tint is raised, the converse is the case. It should be borne in mind, that the elongation of a mineral is positive if the direction of elongation coincides with the vibration direction n_x , and negative, if the direction of elongation coincides with n_z . This can be checked by rotating the mineral through 90° and then observing the reverse phenomena (Fig. 29).

If the mineral's double refraction is strong, the gypsum plate cannot be used to determine the position of the axes n_x and n_z , and another compensator, the quartz wedge, must be used.

When the sharp edge of the wedge is gradually inserted into the slot, we obtain a succession of all the interference colours from dark grey at the thin end to fourth order colours at the thick end. To determine the character of the elongation by using the quartz wedge, the mineral must be brought into the position of maximum illumination (45° from the extinction position), and then the wedge is gradually inserted so that its axis runs parallel to the length of the mineral and the interference tints of the mineral can be made out to be raised or lowered and the corresponding sign of elongation determined just as in the case of the gypsum.

As for the relationship between the sign of the principal zone of the crystal and the sign of the crystal as a whole, a distinction must be made between uniaxial and biaxial crystals. The axes n_o and n_e occur in any elliptical section of the indicatrix of a uniaxial crystal, and their relative sizes determine the sign of the principal zone and the sign of the crystal. If the crystal is of prismatic habit, then the sign of the principal zone is the sign of the crystal. If the crystal is, however, of tabular habit, then the sign of the principal zone and the sign of the crystal are opposite.

In biaxial crystals, the indicatrix is characterized by the relative sizes, not of two but of three principal axes and the relationship between the sign of the principal zone and the sign of the crystal is more complex. The optical sign of a biaxial crystal is determined in converging polarized light.

Twins under the microscope. Sections of individual crystals extinguish or are illuminated as a whole. Between crossed nicols, the surface of simple twins is divided along the composition plane into two parts, one of which extinguishes when the other is illuminated.

A great number of uniform twin lamellae are seen in sections of polysynthetic twins, and these extinguish as shown in Fig. 30.

If the twin plane and the composition plane coincide (the most usual case), then the section perpendicular to the twin plane will, in both individuals of the twin, give extinction angles which are equal, but positioned in opposite directions relative to the composition face.

If the plane of the section intersects the twinning plane of the crystal at a very small angle, then the two individuals of the twin will overlap one another for a considerable part of their width, and we shall see a narrow band between them, which is optically different from both individuals.

A twinning complex makes it possible to distinguish, for example, orthoclase (simple twins) and quartz (twinning absent) from plagioclase (polysynthetic twinning) and microcline (reticulate twinning) (Fig. 31).

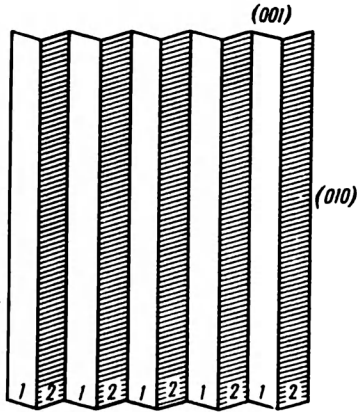


Fig. 30. Polysynthetic twinning

Zoning. Crystals of a substance which gives an isomorphous mixture, for example, crystals of pyroxenes or feldspars, often show zoning under the microscope. This phenomenon is expressed in that their composition progressively (or periodically) changes from the centre to the periphery. There is a parallel change in the optical properties, for example, the colour (tourmaline) and extinction angle (plagioclases).

The phenomenon of zoning can be explained by change in the physical and chemical circumstances during the crystallization of the mineral, as, for example, the higher melting point of the anorthite component (1555°C against 1180°C for albite), which accumulates predominantly in the first stages of the crystallization



Fig. 31. Cross-hatched twinning in microcline

of the isomorphous mixture; as the temperature falls, the later crystallizing zone is enriched in the easier melted albite.

Optical anomalies. When certain minerals are studied under the microscope, optical phenomena are sometimes observed which do not correspond with the symmetry usually attributed to the mineral. Such phenomena are termed optical anomalies. Thus, for example, sections of uniaxial crystals perpendicular to the optic axis should be optically isotropic, but they sometimes show a grey interference tint which is irregularly distributed, or in the form of sectors and bands; a uniaxial crystal is observed to be biaxial, and so on. The cause of an anomaly most often appears, among other causes, to be various inclusions, varying composition (especially in isomorphous mixtures), or the effect of mechanical strains appearing in the process of solidification of the melt, a lowering of the symmetry as a consequence of structural transformation, etc.

4. THE STUDY OF MINERALS IN CONVERGENT LIGHT (CONOSCOPY)

If a short focus Lasaulx lens, or condenser, is introduced between the polarizer and the thin section, then only one ray will proceed parallel to the optic axis of the microscope. The remaining rays of the cone will converge at one point in the centre of the crystal, and then, on emerging, will diverge in all directions, again as a cone.

When a beam of such converging rays is transmitted through a plate of a uniaxial crystal cut perpendicular to the optic axis and between crossed nicols, we shall obtain a dark spot in the centre of the field of view, because the rays passing through the centre are parallel to the optic axis. Concentric circles of dark and bright bands will be observed around this dark spot in monochromatic light. In white light, these bands will show the usual interference colours. Where the path difference is equal to a whole number of wave lengths, extinction is observed to take place along the cross hairs in the ocular. As a result, we see a dark cross intersecting bright rings. This is called an interference figure, and its study in convergent light is termed conoscopy.

The use of convergent light makes it easy to distinguish uniaxial crystals from biaxial crystals, to find the optic sign of crystals, to measure the optic axial angle, and to study dispersion phenomena.

Uniaxial crystals. A section of a uniaxial crystal perpendicular to the optic axis, i.e., a section in which the optic axis of the crystal coincides with the axis of the microscope, will be dark in parallel light between crossed nicols. By inserting the condenser, it becomes possible to study the conoscopic picture just described. It should

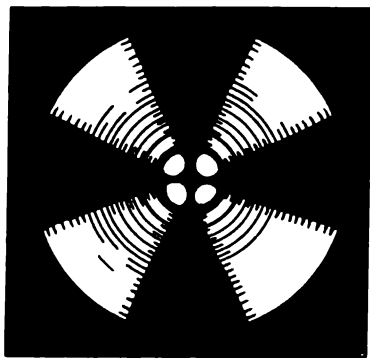


Fig. 32. The interference figure given by a very thick plate of a uniaxial crystal with high birefringence

be noted that the more light and dark rings will be seen and the darker the cross will be, the greater the strength of the double refraction (Fig. 32). The strength of the double refraction of a mineral under investigation can be judged from the number of isochromatic rings when the section has the same thickness, and this is often of importance in determining coloured minerals, whose interference colours are distorted by their natural colouring. Thus, for example, for a 0.03 mm section: quartz ($n_e - n_o = 0.009$) cut perpendicular to the optic axis gives a black cross and the first yellowish ring at the periphery of the field of view; biotite ($n_x - n_z = 0.040$)—two coloured rings, and calcite ($n_o - n_e = 0.172$)—many rings.

In crystals with weak double refraction and in thin preparations, the first ring appears beyond the boundaries of the field of view, and the latter then shows only the dark cross.

An eccentric interference figure is obtained when the section is inclined; when the slice is very oblique, only one branch of the dark cross may be seen or the cross may be absent altogether (Fig. 33). When only one dark band is seen, the figure can be distinguished from a biaxial one by rotating the microscope stage. In the case of a uniaxial crystal, the ends of the branches of the cross will appear in turn, the dark branch of the cross remaining straight, i.e., moving parallel to itself (Fig. 34).

Biaxial crystals. Biaxial crystals are characterized by sections perpendicular to the acute bisectrix. When these sections have their axes parallel to the principal sections of the nicols, the same principles apply as in the case of uniaxial crystals: in white light, they give a system of coloured curves, not circles, however, but like a figure 8; the curves are intersected by a dark cross, made up of a wider band parallel to the optic normal, and a narrower one, parallel to the obtuse bisectrix (Fig. 35).

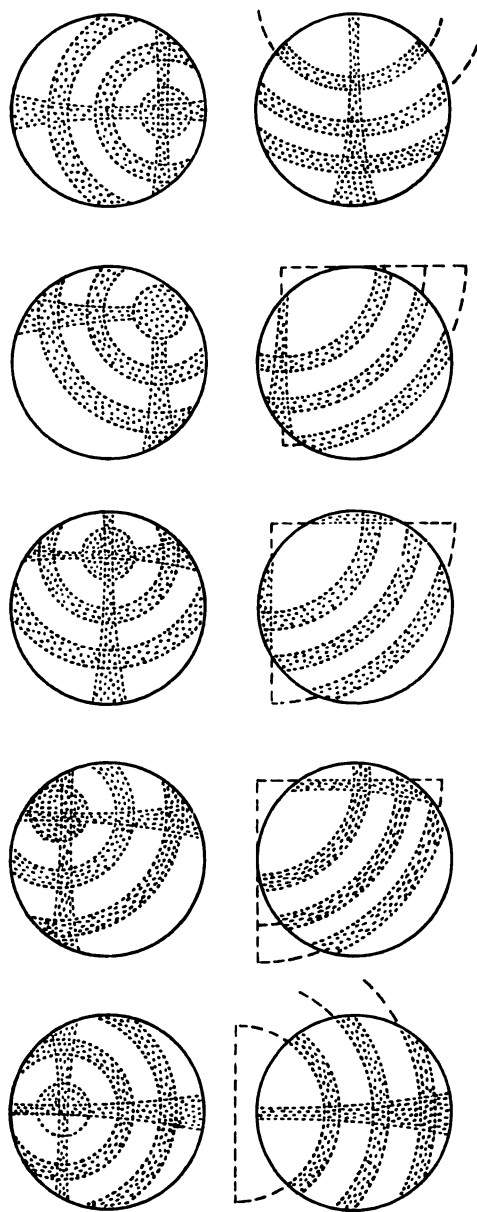


Fig. 33. Interference figures of uniaxial crystals cut obliquely to the optic axis

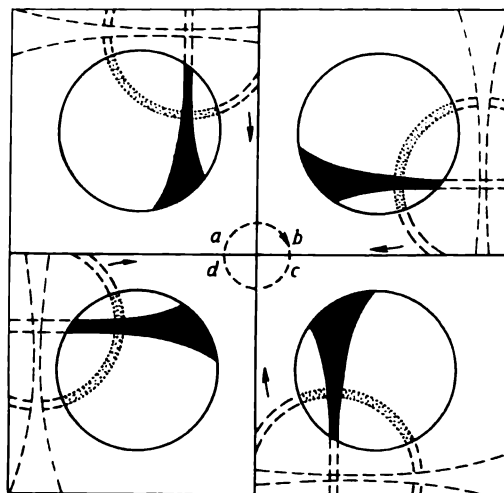


Fig. 34. Very oblique section. Only one isogyre can be seen, and it moves parallel to itself when the microscope stage is rotated

On the latter band, at equal distances from the centre of the cross are set two points surrounded by rings. These are the points of emergence of the optic axes, and the rings surrounding them are analogous to the rings observed around the emergence of the optic axis in the interference figure of uniaxial crystals.

When the microscope stage is rotated, the cross opens up and is replaced by the branches of a hyperbola with the apices of the branches coinciding with the emergence of the optic axes (Fig. 36).

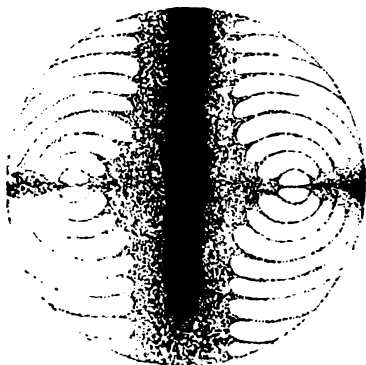


Fig. 35. The interference figure of a biaxial crystal in a section perpendicular to the acute bisectrix. The axes of the section are parallel to the principal section of the nicols

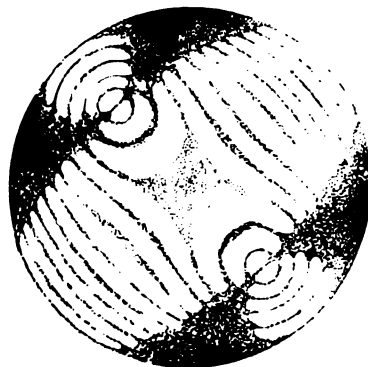


Fig. 36. The same section in the 45° position

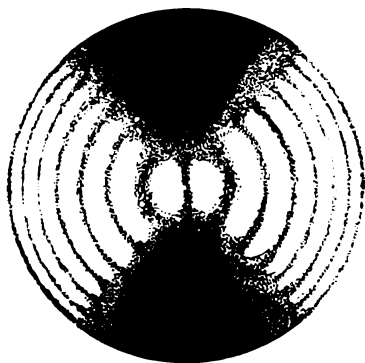


Fig. 37. The interference figure of a biaxial crystal in a section perpendicular to one of the optic axes, $2V = 90^\circ$

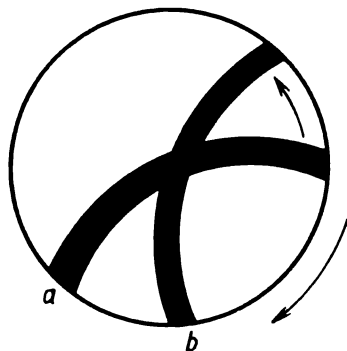


Fig. 38. The change in the position of the same figure when the microscope stage is rotating

The hyperbola, like the cross, is the geometric locus of points in which the vibration directions coincide with the principal sections of the nicols.

The dark cross which opens up into a hyperbola when the stage is rotated is always seen in the conoscopic in sections perpendicular to the acute bisectrix.

If the preparations are not thick enough, or the double refraction not sufficiently high, the coloured curves may be absent.

Sections perpendicular to one of the optic axes give half a figure perpendicular to the acute bisectrix, and this shows coloured concentric rings intersected by one dark band (Fig. 37), which bends when the stage is rotated (in distinction to uniaxial crystals, in which the band remains straight) and moves in a direction opposite to the rotation of the stage (Fig. 38). Its convex side is always turned towards the acute bisectrix.

Sections of biaxial crystals, not giving any emergence of the optic axes in convergent light, and those sections perpendicular to the obtuse bisectrix and the optic normal n_m , give an indistinct cross in the conoscopic, which rapidly passes into a hyperbola when the stage is rotated.

5. FINDING ORIENTED SECTIONS

A section of a *uniaxial mineral* parallel to the optic axis (Fig. 39,3) shows the highest interference colours in parallel light, these colours diminishing as the section is inclining to the optic axis (Fig. 39,2), and at an inclination of 90° , i.e., perpendicular to the optic axis (Fig. 39,1), the section appears isotropic, i.e., it is black.

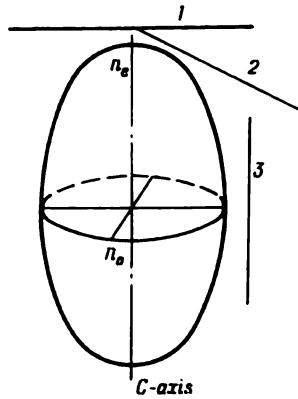


Fig. 39. Disposition of an oriented section in a uniaxial crystal

Sections of *biaxial crystals* parallel to the optic axial plane, i.e., perpendicular to the optic normal, show the highest interference colours. A section perpendicular to the acute bisectrix (Fig. 40,1) will show the lower interference colours, the smaller the optic axial angle. The interference tint in a section perpendicular to the obtuse bisectrix (Fig. 40,3) is intermediate; it is the nearer the highest colour, the greater the obtuse optic axial angle. Sections perpendicular to one or other of the optic axes (Fig. 40,2) are not illuminated at all.

The optical character (sign) of uniaxial and biaxial minerals is determined in sections which give an emergence of an optic axis by means of the gypsum plate or quartz wedge. A uniaxial crystal can be regarded as the particular case of a biaxial crystal, when $2V = 0$, and therefore the phenomena which serve to determine the optical character in convergent light will be analogous for both uniaxial and biaxial minerals.

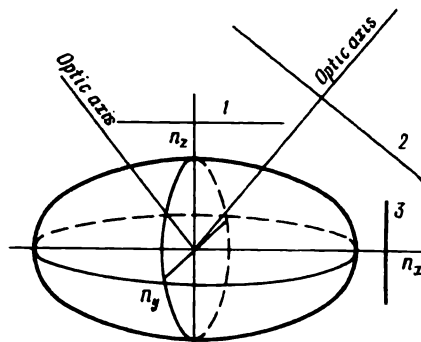


Fig. 40. Disposition of an oriented section in a biaxial crystal

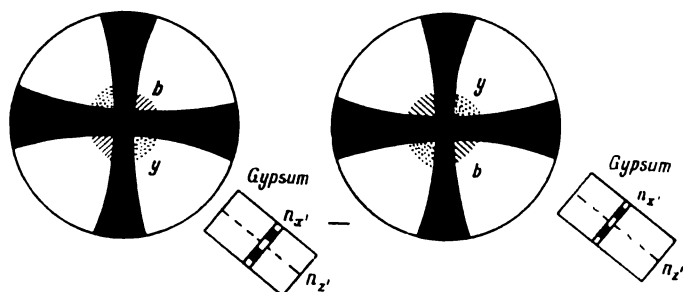


Fig. 41. Determination of the sign of a uniaxial mineral using a gypsum plate and the change of colour in the interference figure when the plate is inserted

If, having obtained a conoscopic picture of a uniaxial crystal, a gypsum plate or quartz wedge is inserted into a diagonal slot in the microscope tube, then in the quadrants arranged along the length of the plate, there will be obtained a subtraction of the path differences of the crystal and plate in the case of a positive crystal, and an addition of the path differences in the case of a negative crystal, because axes of the same designation coincide. In quadrants arranged across the width of the gypsum plate (or quartz wedge) the resulting phenomena are directly opposite. When the gypsum plate is inserted, a summation (negative crystal) corresponds to a blue colour, while a subtraction (positive crystal) corresponds to a yellow tint within the confines of the first ring of the interference figure (Fig. 41). When the quartz wedge is pushed in, the coloured rings in a negative crystal contract in the direction of movement of the wedge, as if they were running towards the centre of the cross, towards the point of emergence of an optic axis of the crystal; in

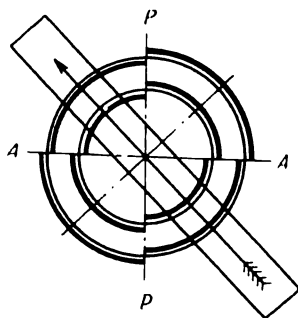


Fig. 42. The effect of inserting a quartz wedge into the interference figure of a uniaxial negative crystal

the case of an optically positive crystal, on the contrary, a widening and moving apart of the rings is seen, running off from the centre in the same direction as the wedge is moving (Fig. 42).

To determine the optical sign of a biaxial crystal in the conoscope, use is made of all sections which give the emergence of at least one optic axis; and it is desirable, but by no means indispensable, that the optic axis emerges, if possible, in the centre of the field of view. In parallel rays, such sections are recognizable by their comparatively weak double refraction.

To determine the sign of a biaxial mineral, the microscope stage is turned until the dark hyperbola (or one of its arms) is situated in the field of view in a direction perpendicular to the movement of the gypsum plate or quartz wedge, and the wedge (or plate) is then inserted.

The convex side of the hyperbola arm faces the acute bisectrix, while the concave side faces the obtuse bisectrix. If the crystal in question is negative, then, on the convex side of the arm, are seen phenomena corresponding to the case of a positive uniaxial crystal (yellow patches for gypsum and a movement away of the lemniscates from the optic axis towards the acute bisectrix for the quartz wedge), while on the concave side phenomena corresponding to the case of a negative uniaxial crystal (blue patches with gypsum; a concentration of rings from the obtuse bisectrix to the optic axis for the quartz wedge).

So, if a blue colour appears on the convex side of the isogyre when a gypsum plate is inserted, and a yellow colour on the concave side, then the mineral is positive, and vice versa (Fig. 43).

Actually, the axis n_y (Fig. 44) is always arranged perpendicular to the optic axial plane. If, then, the ray in question passes between the optic axis and the acute bisectrix (on the convex side of the isogyre) of a positive crystal, then its vibrations parallel to the optic axial plane will proceed along n_z , while for the ray emerging between the axis and the obtuse bisectrix (on the concave side of the isogyre), n_x is situated along the same direction.

Measurement of the optic axial angle. The optic axial angle is a very characteristic constant for a great number of minerals. A distinction has to be made between the real angle $2V$, formed by the optic axes within the mineral, and the apparent angle $2E$, which is formed by them on emergence from the mineral, and seen by us in the conoscope (Fig. 45). The ratio of the two angles is $\sin V = \sin E/n_y$, where n_y is the mean refractive index of the given mineral. In section, the optic axial angle can be determined indirectly by using a micrometer scale to measure the distance between the emerging optic axes in converging light ($2D$). Mallard's formula is then employed

$$\sin E = DK$$

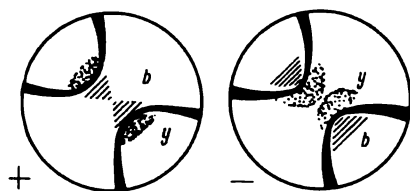


Fig. 43. Change of colour on the different sides of the hyperbola for a biaxial crystal

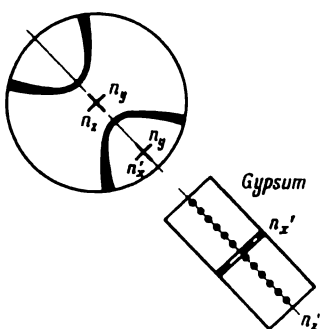


Fig. 44. Explanation of the appearance of colours with the gypsum plate on different sides of the isogyre of biaxial crystals

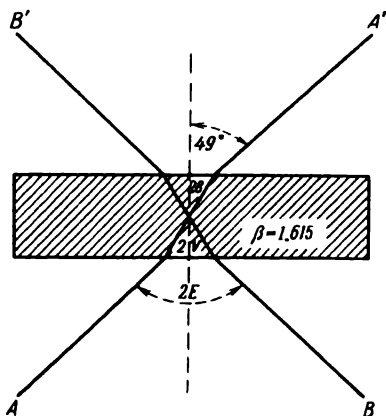


Fig. 45. Real ($2V$) and apparent ($2E$) optic axial angles

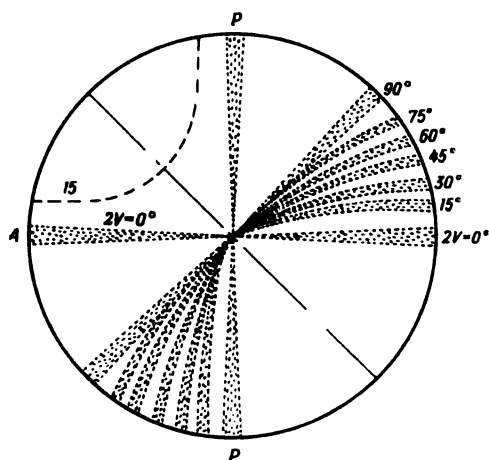


Fig. 46. Approximate determination of $2V$

where D = linear distance between the axis of the microscope and the apex of the hyperbola;

K = certain constant, calculated for the given optical system.

For an approximate estimation of the value of $2V$, the curvature of the isogyre in sections perpendicular to the optic axis can be determined (Fig. 46). If the sides of the isogyre are bent at an angle near 90° , $2V$ will be small or even near zero; when the isogyre is straight, $2V$ is equal to 90° ; intermediate forms correspond to medium values of $2V$.

The phenomenon of dispersion. Dispersion is the name given to the change of any optical constant in connection with a change in the wave length of light. The character of the dispersion depends on the symmetry of the crystal; thus, in uniaxial crystals, only the value of the refractive index can change, and the only dispersion that occurs is that of the refractive index.

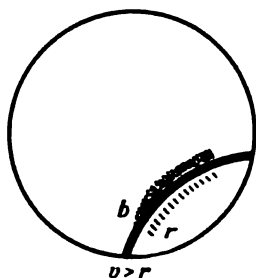


Fig. 47. Dispersion of the optic axes of a biaxial crystal (the optic axial angle is greater for blue light than for red)

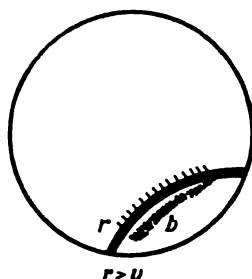


Fig. 48. Dispersion of the optic axes of a biaxial crystal (the optic axial angle is greater for red light than for blue)

In orthorhombic crystals, the position of the indicatrix is strictly fixed, therefore only the refractive indices can change, but as a result of dependence on these, the optic axial angle also changes and is different for light of different wave lengths; therefore, dispersion of the optic axial angle takes place. Under the microscope, this can be seen from the red or violet shade with which an isogyre may be tinted.

This is written as $v > r$, if the optic axial angle is greater for blue light than for red (Fig. 47), and as $r > v$, if the position is the reverse (Fig. 48).

In monoclinic crystals, as well as dispersion of the optic axial angle, there can occur dispersion in the position of the bisectrices or dispersion in the position of the optic normal, depending on which axis of the indicatrix coincides with the b axis of the crystal, along which no dispersion can take place.

In triclinic crystals, there can also be asymmetric dispersion, expressed as a change in the position of all the axes of the indicatrix.

It should be pointed out that dispersion is insignificant in the majority of crystalline substances and is discerned with difficulty.

6. DETERMINATION OF CRYSTAL SYSTEM FROM OPTICAL PROPERTIES

In the majority of cases, the system to which an unknown mineral belongs can be determined from the extinction angle and double refraction measured in several sections, and from whether it is uniaxial or biaxial.

Cubic system: simple refraction; all sections are isotropic, and, like glass, are not illuminated between crossed nicols.

Tetragonal, trigonal and hexagonal systems: double refraction and uniaxiality are distinguishing features; straight extinction of elongated sections of crystals of prismatic habit, and in certain transverse sections in crystals of tabular habit. It is difficult to distinguish between these systems by optical properties; the determination is usually facilitated by the shape of the section or the character of the cleavage (a triangular section is typical of trigonal crystals, a square section of tetragonal, and six-sided section of hexagonal crystals).

Orthorhombic system: crystals are doubly refracting and biaxial; straight extinction is characteristic in elongated sections of crystals of prismatic habit and in sections parallel to the pinacoids or cleavages.

Monoclinic system: crystals are biaxial and have one direction in which there is straight extinction. In all other sections, the extinction is oblique. Certain crystals can be distinguished according to whether the intersection of the principal cleavages (and the

direction of elongation of the crystals) lies in the symmetry plane or perpendicular to it. In the first case, the elongated sections can give some extinction angle between 0° and the maximum value characteristic for the mineral species or variety; in the second case (for example, epidote and wollastonite), the elongated sections give straight extinction. The first case is more frequent.

Triclinic system: crystals are biaxial, but there is no section giving straight extinction.

7. ADDITIONAL PRACTICAL INSTRUCTIONS IN CRYSTAL OPTICS

1. Even the most perfect cleavage cannot be seen in a section of a mineral when the cleavage planes make a low angle with the plane of the section.

2. In an isotropic section and in sections near it, even a pleochroic mineral will not show its pleochroism.

3. The double refraction of a mineral is characterized by the highest interference tint (in the principal section) for a given thickness of slice. In other sections, it can have a lower tint until it is black (in an isotropic section).

In slides of normal thickness, quartz and plagioclase should have an interference tint not higher than yellowish white, and orthoclase not higher than pale grey.

A quartz wedge is a most useful means of studying interference colours of the different orders, as it usually gives the colours of the first four orders; colours higher than the fourth order then approximate to white.

4. In any anisotropic section the relative refraction (and that means the shagreen surface and the Becke line) must be studied for both of its principal axes (with darkness between crossed nicols as control).

5. The following tentative grouping of minerals in regard to their boundaries, relief and surface has been proposed by V. Lodochnikov and will be found useful:

Group 1. Boundaries and relief inconspicuous; if the mineral is fresh, the surface is almost not different from that of the Canada balsam at the edge of the thin section or in the cavities of the latter filled with the Canada balsam; this corresponds to a value of n from 1.53 to 1.56. To this group belong the majority of plagioclases, nepheline, quartz, cordierite, and basic and intermediate volcanic glasses.

Group 2. Boundary and relief are weak, and the surface slightly rough, $n = 1.48$ to 1.52 (negative relief and surface)*. Such are

* Negative relief or a negative shagreen surface is best recognized by the Becke line.

acid and intermediate volcanic glasses, the feldspathoids with the exception of nepheline, cristobalite, zeolites, the potash-soda feldspars.

Minerals with $n = 1.57$ to 1.60 are similar to those in group 2, but possess positive relief and surface. These comprise basic volcanic glasses, bytownite, anorthite, iron cordierite, talc, most chlorites, sericite and muscovite.

Group 3. Boundaries and relief are clear, the surface is distinct and shagreen, $n = 1.41$ to 1.47 (negative relief and shagreen surface). Includes opal, fluorite and tridymite. Minerals with $n = 1.61$ to 1.65 (positive relief and surface) are similar to those in group 3, i.e., apatite, topaz, andalusite.

Group 4. Boundaries and relief are sharp as well as the shagreen surface, $n = 1.66$ to 1.78 . Includes olivine, pyroxenes, epidote, vesuvianite and garnets.

Group 5. Boundaries are very dark and rough, the relief and shagreen surface are extremely sharp. $n > 1.78$. Some garnets, sphene, zircon, rutile, brookite, anatase.

If the boundary cannot be discerned in transmitted light, the analyzer should be introduced, the boundary found and brought into the centre of the field of view; the general picture borne in mind and the analyzer then withdrawn. An attempt should be made to make out the boundary in transmitted light.

Directions for the observation of the Becke line:

1. An intermediate magnification should be used, for example, an objective with a magnification of $\times 40$.

2. The microscope mirror must be lowered as far as possible, as should also the illuminating lens together with the polarizer in a direction towards the mirror.

3. The illumination is adjusted by gradually closing the aperture of the diaphragm (above or below the polarizer); it will be seen that excessive use of the diaphragm weakens the effect.

8. THE PRINCIPLES OF THE FEDOROV METHOD

Work with the polarizing microscope in the investigation of some mineral grain is confined to a random plane along which the section happens to have been made. Consequently, in order to find the identifying planes and directions and to complete the optical properties of a given crystal, observations must be made of many grains of the mineral in question, and this protracts and encumbers the investigations. This inconvenience can be removed to a certain extent by using the *Fedorov universal stage*, which is often and widely employed in studies of crystal optics. Its advantage consists in the fact that it allows the preparation (slice) to be rotated around sev-

eral axes (four or five) which intersect at one point in the centre of the universal stage.

The model mostly used is the four axial stage represented in Fig. 49.

Since the Fedorov stage can be rotated in all directions, it is very useful in studying the position of cleavage planes, and of the elements of the optical ellipsoid and of twinning elements. Therefore, use of the stage makes possible an explanation of the interconnection of crystallographic and optical directions, for instance, the optical orientation of twin formations or the connection between extinction angles and cleavage.

The links have been well studied in the case of the plagioclases, it having been shown that they depend directly on their chemical composition. Similar observations are constantly being made for certain other minerals.

The Fedorov stage finds great application in carrying out various structural investigations (petrotectonics) requiring the greatest precision in the determination of the arrangement and distribution of minerals in rocks.

The convenience of the Fedorov method is that the determination of the optical orientation of any mineral is a geometric operation, which is reduced to the plotting of the axes and planes of symmetry of the optic indicatrix (n_x , n_y and n_z) and of any crystallographic directions, for example, cleavage and twin planes, on to a spherical

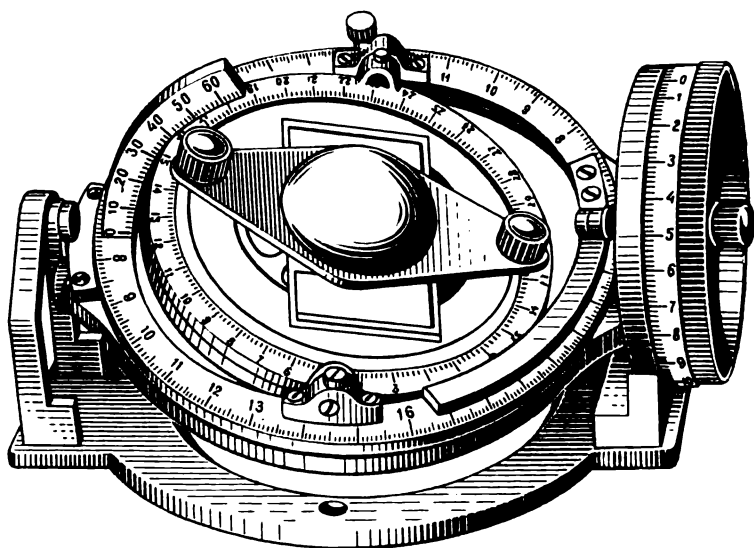


Fig. 49. The Fedorov stage with four axes of rotation

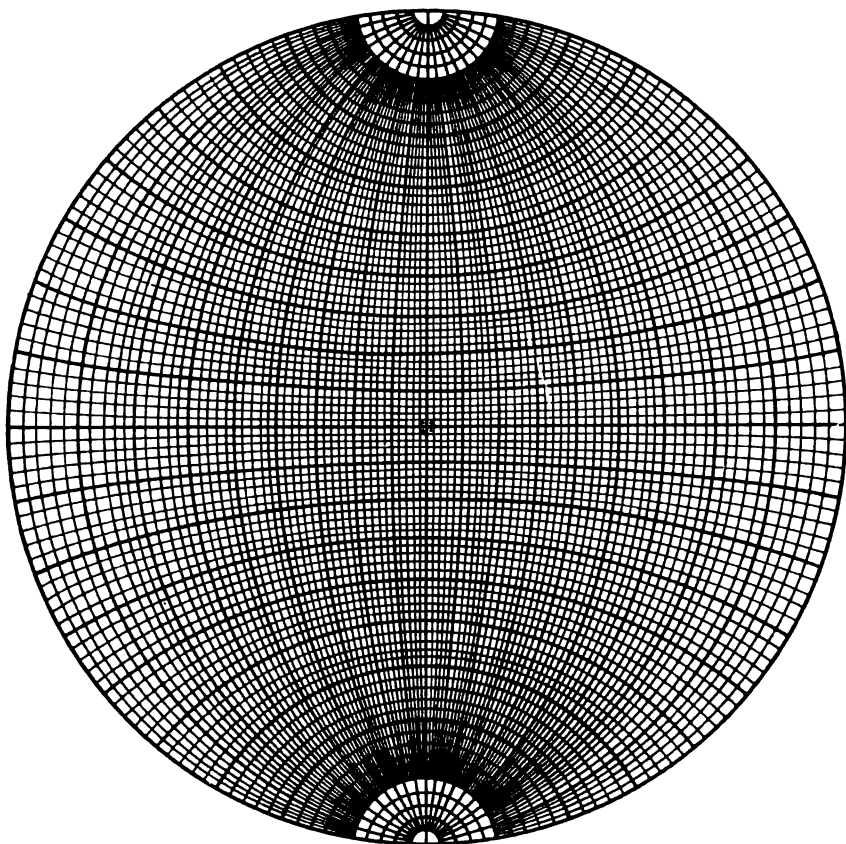


Fig. 50. The Wolff net (reduced approximately 3 times; the standard net is 20 cm in diameter)

(stereographic) projection. The angles between them determine the orientation of the mineral completely. The most convenient way of doing this is the use of the *Wolff net*, wherein meridians and parallel circles are stereographically projected in the plane of the circle at two-degree intervals (Fig. 50). Drawings using the Wolff net are made on tracing paper laid over it.

The Fedorov stage was subsequently improved by adding the fifth axis. This stage can measure directly the coordinates of the crystallographic directions, the orientation and composition of any rock-forming mineral without subsidiary graphical constructions. The proposer of this method, Academician A. Zavaritskii, called it the "twin theodolite method".

9. THE STUDY OF ROCKS IN THIN SECTION

The study of rocks in thin section begins with low magnification. If it is desirable to study some object at high magnification, it should be brought to the centre of the field of view before the objective is changed.

The slice should always be first examined with the analyzer out. Many rock-forming minerals can be determined in most cases from their outlines, relief, cleavage, inclusions, alteration products, etc., without the use of polarized light. Opaque minerals are always examined in reflected light.

To investigate the pleochroism of any coloured mineral, the lower nicol, or the microscope stage, is rotated. The first method is preferable when testing weak pleochroism. Oblique illumination is sometimes used. For example, the extraordinarily fine needles of apatite in some rocks become visible only in such circumstances. Oblique illumination can be obtained by placing a short-focus convex lens under the microscope stage, with the central portion of the lens covered by a disk of black paper.

For high magnifications, the focus is markedly different for the upper and lower surfaces of the section. In order to investigate a body occupying the whole thickness of the slice, the microscope tube must be gradually moved, thus bringing different levels in the slice successively into focus.

The determination of minerals is only a part of a petrographical investigation, for after it begins the study of the rock itself. To this end, the mutual relations of the minerals must be observed, as well as the order of crystallization, intergrowths, inclusions, decomposition products, pseudomorphs, and so on.

CHAPTER VII

Rock-forming minerals

From a total of more than 3000 mineral species, a little more than 100 are commonly found in rocks. These minerals have come to be associated as the *rock-forming minerals*.

In petrography, the polarizing microscope is the main tool of the investigator, and so he needs a good knowledge of the optical properties of these minerals and the ability to identify them microscopically. In a number of cases, however, it is not enough merely to recognize the mineral: plagioclases, for example, are isomorphous mixtures of the albite and anorthite end-members. Consequently, the task will be to use the optical properties of a mineral as a means of determining its composition, or, in other words, to find out its plagioclase number. The same applies to a series of other minerals.

Some rock-forming minerals are the main constituents of igneous rocks (feldspars, hornblendes, pyroxenes, micas, quartz) to an extent measured in per cent or tens of per cent. These are called the essential (main) rock-forming minerals. There are other minerals, which although found in almost every rock, nevertheless occur in quite small amounts, not more than 1%, for example, magnetite, pyrite, titanite, zircon, apatite. It was at first thought that these were accidental additions to the rock, and so they were called *accessory minerals*. It has since been established that these minerals are a systematic part of the composition of the rock, and that there is as much right to speak of zircon or titanite granites as of microcline granites or of hornblende gabbro.

Different conditions of rock genesis give different mineral assemblages.

V. Lodochnikov, a most eminent petrologist, has pointed out that the rock-forming minerals are the stones out of which rock structures are built, and that the minerals are physico-chemical units, the rocks being physico-chemical systems made up of these units.

Temperature, pressure, concentration of the components and medium are the main parameters controlling the course of a physico-chemical process involved in rock formation, and a change in even one of these parameters can cause a change in the composition of the rock. Accordingly, the conditions under which rocks were created can be established when the data from detailed rock investigations are available. This is particularly effective in the connection between petrology and the genesis and paragenesis of minerals, and between petrology and investigations of useful mineral deposits.

Mineral assemblages are a most important diagnostic feature in the study of a rock, being the outcome of its chemistry (and consequently of the physico-chemical circumstances under which they crystallized). Every geological process is characterized by its own particular physico-chemical conditions, and by its own mineral paragenesis. Thus, feldspars, or apatite, are formed only in magmatic medium; volcanic glass is typical of effusives; quartz, calcite, barytes are usually vein minerals of hydrothermal formation. In the course of weathering, the soluble components of magmatic and any other rocks are borne away to the waters of oceans and lakes. When lakes dry up, or there is a local increase in the concentration of some salt or other in different areas of the ocean, conditions can be created for their precipitation as a sediment, and so for the formation of sedimentary rocks typified by various carbonates (calcite, siderite, dolomite), soluble salts (halite, sylvite and others), gypsum, carbonaceous materials and various hydrated metal oxides. Earlier formed minerals which are stable enough to exist at the surface: quartz, clays, mica, zircon and many others, may occur in the sediment as mechanical additions, but do not, of course, define the conditions under which the new rock was precipitated.

The conditions of mineral formation are especially complex where a rock is subjected to unaccustomed high temperatures and pressures, or to the introduction of active solutions.

In all cases of metamorphic rock formation, there is a specific association of newly formed minerals which is characteristic of these rocks only. Here, along with minerals widely distributed in magmatic rocks (feldspars, quartz, hornblende, pyroxene), we find minerals that are formed only in metamorphic processes.

Technology is comparable with metamorphism in that they both produce entirely new minerals, but since technological processes involve high temperatures and other conditions not encountered in nature, and because the composition of the raw materials must be specific, these new minerals seldom occur naturally. They include mullite—the mineral found in porcelain and refractories, the cement minerals—alite and belite, the abrasive mineral—corundum, and the slag minerals—fayalite, cristobalite and tridymite.

The silicates are the most important of the rock-forming minerals, and their structures are fairly well known now. Silicon is the acid

element in them. Every atom of silicon in the silicates is surrounded by four oxygen atoms arranged at the corners of a tetrahedron. If these tetrahedral groups are not linked to one another by one or more of the oxygen atoms forming the tetrahedron, the formula of such a substance must include one or more SiO_4 groups as, for example, in the case of zircon (ZrSiO_4) and forsterite (Mg_2SiO_4).

If the crystal structure contains tetrahedral groups joined together in pairs, the formula should contain Si_2O_7 (for example, âkermanite $\text{Ca}_2\text{MgSi}_2\text{O}_7$). If they are joined into a group of three, forming a trigonal ring, then the formula will include the Si_3O_9 group. A group of four tetrahedra forms a tetragonal ring and the formula contains Si_4O_{12} . Six SiO_4 groups form a hexagonal Si_6O_{18} ring (for example, beryl $\text{Be}_3\text{Al}_2\text{Si}_6\text{O}_{18}$). When the tetrahedra form a continuous chain, as in the pyroxenes, the formula should include the SiO_3 group (or multiples of it). In the amphiboles, the tetrahedra are joined in the form of double chains or bands with a formula that includes Si_4O_{11} . We find sheets (or layers) of tetrahedra in the micas, and their formula contains the Si_4O_{10} group (or multiples of it). If the tetrahedral groups form a continuous, unbroken, three-dimensional network, as in quartz, the formula will then include SiO_2 . Such a continuous three-dimensional bonding consisting entirely of SiO_2 groups can be built up in various ways (quartz, tridymite and cristobalite).

The Si atoms in the centre of some tetrahedral groups can be replaced by atoms of other elements, especially by Al. If one fourth of the Si atoms in a continuous three-dimensional lattice are replaced by Al atoms, the arrangement can be represented as $(\text{Si}_3\text{Al})\text{O}_8$. Such a grouping of atoms is unstable, since one oxygen valency remains free. But if this valency is satisfied by an atom of K or Na, these substances will become stable and crystallize as orthoclase $\text{K}(\text{AlSi}_3)\text{O}_8$ or albite $\text{Na}(\text{AlSi}_3)\text{O}_8$. If one third of the Si atoms is replaced by Al, the formula should contain $(\text{Si}_2\text{Al})\text{O}_6$ or multiples of this expression: in the case of analcite $\text{Na}(\text{Si}_2\text{Al})\text{O}_6 \cdot \text{H}_2\text{O}$. When half of the Si atoms are replaced by Al atoms, the formula changes to $(\text{SiAl})\text{O}_4$ or multiples of it, e.g., in nepheline $\text{Na}(\text{SiAl})\text{O}_4$ and in anorthite $\text{Ca}(\text{Si}_2\text{Al}_2)\text{O}_8$. Atoms of Al can replace Si atoms not only in continuous three-dimensional lattices, but in other silicate structures as well. Combinations of several types of structural elements are sometimes encountered.

The main features of the atomic structures of some silicates are presented in Table 2.

Authors of textbooks adhere to different groupings when describing rock-forming minerals. V. Luchitskii, for instance, divides all minerals into opaque and transparent. The transparent minerals are in their turn divided into isotropic and anisotropic (uniaxial and biaxial). Within these groups the arrangement is according to increasing refractive index. V. Lodochnikov divides all minerals

Table 2

Main Features of the Atomic Structure of Certain Silicates

(Si, Al):O	Acid radical	Grouping of radicals	Examples	
			formula	mineral
12:48	SiO ₄	Separate tetrahedra	Mg ₂ SiO ₄	Forsterite
12:42	Si ₂ O ₇	Double tetrahedra	Ca ₂ MgSi ₂ O ₇	Åkermanite
12:36	SiO ₃	Trigonal rings	BaTiSi ₃ O ₉	Benitoite
12:36	SiO ₃	Square rings	Si ₄ O ₁₂	—
12:36	SiO ₃	Hexagonal rings	Be ₂ Al ₂ Si ₆ O ₁₈	Beryl
12:36	SiO ₃	Chains	MgSiO ₃	Enstatite
12:33	Si ₄ O ₁₁	Bands	(OH) ₂ Mg ₇ (Si ₄ O ₁₁) ₂	Anthophyllite
12:30	Si ₄ O ₁₀	Layers	(OH) ₂ Mg ₃ Si ₄ O ₁₀	Talc
12:30	Si ₄ O ₁₀	Layers	KF·2Ca ₂ Si ₄ O ₁₀ ·8H ₂ O	Apophyllite
12:24	SiO ₂	Framework	SiO ₂	Quartz
12:24	(Si ₃ Al)O ₈	Framework	K(Si ₃ Al)O ₈	Orthoclase
12:24	(Si ₂ Al)O ₆	Framework	Na(Si ₂ Al)O ₆ ·H ₂ O	Analcite
12:24	(SiAl)O ₄	Framework	Na(SiAl)O ₄	Nepheline

into colourless and coloured, and of the colourless separates seven groups according to features depending on the refractive index. In view of the brevity and special aims of the present course, there are several departures from the generally accepted method of classifying minerals. Abbreviated data for the rock-forming minerals are given below, not according to physical (optical) groups, but according to their genetic associations; it has not been possible to adhere to this principle completely in view of the variety in the conditions of occurrence and in the properties of the minerals even within the limits of any one group.

The order adopted in describing the minerals can be presented in the following form.

A. MINERALS OF MAGMATIC ROCKS

1. Most Widespread

(a) *Light or pale*. 1. Feldspars: potash-soda (sanidine, orthoclase, microcline, anorthoclase, perthite); plagioclases (albite, oligoclase, andesine, labradorite, bytownite, anorthite). 2. Quartz. 3. Feldspatoids (feldspar substitutes): leucite, nepheline, sodalite, hauyne, nosean, analcite.

(b) *Coloured* (ferromagnesian). 1. Micas. 2. Amphiboles and pyroxenes. 3. Olivine.

II. Secondary

- (a) Accessory: tourmaline, titanite, apatite, zircon, magnetite.
- (b) Secondary: calcite, zeolites, chlorite, talc, epidote.

B. MINERALS OF SEDIMENTARY ROCKS

Opal, calcite, kaolin, dolomite, gypsum, fluorite, limonite, pyrite.

C. MINERALS OF METAMORPHIC ROCKS

Garnet, staurolite, andalusite, sillimanite, cordierite, kyanite.

D. THE MOST IMPORTANT MINERALS IN TECHNOLOGY

Mullite, devitrite, alite, belite, monticellite; cristobalite and tridymite; corundum.

1. THE MOST PREVALENT MINERALS IN MAGMATIC ROCKS

Light-Coloured Minerals

Feldspars are the most widespread mineral group in nature. They comprise about 60% of the earth's crust, and are the main constituent of most magmatic and metamorphic rocks. In composition they are complex framework aluminosilicates of potassium, sodium and calcium, more rarely of barium and sometimes contain some strontium, rubidium or caesium. Feldspars with the first three bases show a marked predominance, so that these K-, Na- and Ca-feldspars are the rock-forming feldspars.

The group of feldspars is united by their crystallographic properties. Although they crystallize in two systems, the monoclinic and triclinic, they form crystals which are very near each other in form, being usually bounded by two strongly developed pinacoids (001) and (010) and a weakly developed prism (110). There is a perfect cleavage parallel to both pinacoids, and since the angle between them is nearly 90° (or equal to 90° in the monoclinic varieties), we obtain little rectangular plates or bars, when cleavage fragments are split off.

Although the pure forms of the potash, soda or calcic feldspars are found in nature, they are comparatively rare. The pure potash molecule KAlSi_3O_8 is known under the name of orthoclase, the soda $\text{NaAlSi}_3\text{O}_8$ of albite, and the calcic $\text{CaAl}_2\text{Si}_2\text{O}_8$ of anorthite. Isomorphous mixtures of these three components are encountered much more frequently, but their miscibility is far from being the same. More than 1500 chemical analyses of natural feldspars are

presented in the form of a diagram in Fig. 51. Each point inside the triangular diagram represents one analysis. The corners of the triangle represent the pure components. The nearer a point is situated to one of the angles, the more of that component is contained in the feldspar represented by the point.

When the diagram is examined, the majority of points can be seen to be concentrated either in the vicinity of the lower side of the triangle (the albite-anorthite side), or near its upper apex (orthoclase). Points are completely absent from the right-hand side of the diagram between anorthite and orthoclase, but on the left-hand side, between albite and orthoclase, they are fairly plentiful. In other words, the feldspars most often encountered are mixtures of the albite and anorthite molecules with a small addition (up to 10%) of orthoclase molecules, and orthoclase molecules with a small addition (up to 10%) of anorthite or a little more (up to 35-40%) of albite.

Feldspars of the albite-anorthite series are united together under the name of plagioclases.

Microscopic investigation of feldspars whose analyses show more than a 30% content of albite in orthoclase, and more than 30% orthoclase in albite, shows that these feldspars belong to what is known as perthite, the main part of a crystal of which is formed of orthoclase, in which albite occurs as inclusions and intergrowths (Fig. 52). At present, the cause of this is ascribed to the fact that mixtures on the right-hand portion of the diagram are

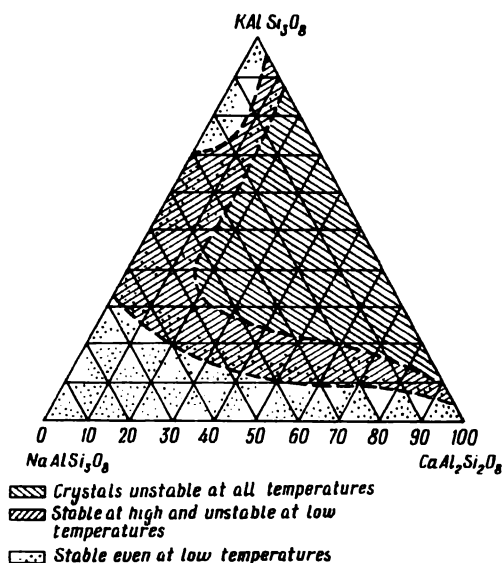


Fig. 51. The stability relationships of feldspars

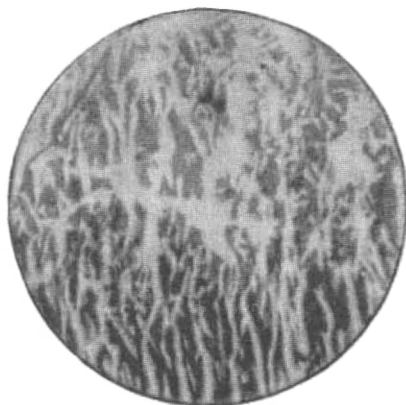


Fig. 52. Microperthite

unstable, and if a magmatic melt has such a composition, two feldspars will crystallize from it simultaneously, forming orthoclase and plagioclase.

Compositions capable of crystallizing in the form of homogeneous crystals at the freezing point of the magma are represented with a special shading on the diagram (see Fig. 51), but with further solidification they prove to be unstable, and already in the solid form they dissociate into orthoclase and perthitic intergrowths of plagioclase (usually albite). The feldspars, whose composition lies in the unshaded portions of the diagram, crystallize freely out of the magma, and preserve their original homogeneity.

The refractive index of feldspars varies from 1.518 to 1.520 for potash feldspar, 1.530 for albite to 1.580 for anorthite, and consequently, in thin section, cemented with Canada balsam with $n = 1.537$, a feldspar will not show any relief, or the relief will be weak.

The double refraction of feldspar is usually 0.006; more rarely, it reaches 0.012; therefore, in a section of normal thickness it always possesses either a grey, or a weak yellow tint of the first order.

Average refraction, weak double refraction together with the characteristic external faces and distinct cleavage, make it easy to distinguish the feldspars from all other minerals. Simple and polysynthetic twinning is easily discernible under the microscope and is very characteristic of feldspars.

Potash feldspars crystallize in two systems, monoclinic and triclinic.

These differences have been explained as due to the arrangement of the aluminium and silicon ions in the crystal lattice. If these ions are distributed throughout the lattice as a whole, but without any order, and occupying any place appropriate to them, the lattice obtained will be of high symmetry (disorder). During the post-

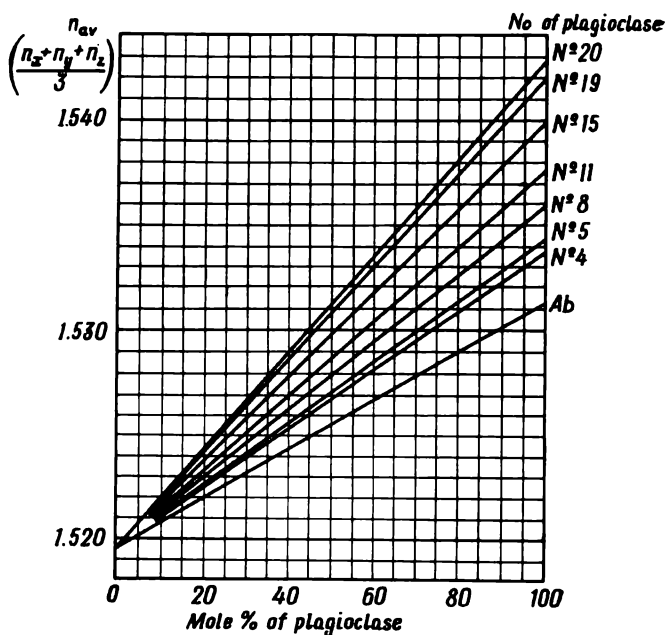


Fig. 53. The dependence of the average refractive index of potash feldspar on the amount of plagioclase it contains (after D. Belyankin)

genetical existence of a rock, an "ordering" of the structure can take place, aluminium and silicon occupying strictly defined positions in the feldspar lattice, as a result of which the symmetry of the crystal is lowered, and lowered the more, the greater the ordering.

Sanidine is the least ordered feldspar, and microcline the most ordered. According to this, sanidine and orthoclase are monoclinic feldspars, and microcline and anorthoclase are triclinic. They can all be distinguished under the microscope.

Microcline is most easily distinguished from the others; it is twinned on two laws, the albite and the microcline, giving indivisible portions which intersect each other almost at right angles and which show, as a result, the very clear and distinctive microcline cross-hatching (see Fig. 31).

Sanidine is usually water-clear; the small optic axial angle ($30-40^\circ$) is its specific feature. Both optic axes are seen to emerge in a section perpendicular to the acute bisectrix, in convergent light, and the isogyre is strongly curved in a section perpendicular to one optic axis.

Anorthoclase has an average optic axial angle ($50-60^\circ$), and consequently the isogyres are not very curved, but are completely shown. Anorthoclase can be distinguished more precisely from orthoclase by determining its triclinic symmetry on the Fedorov stage.

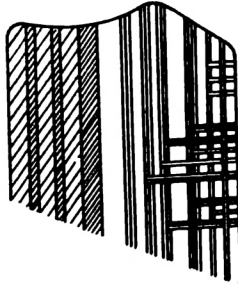


Fig. 54. Complex twinning in a plagioclase crystal, $\times 14$. Crossed nicols

Orthoclase is characterized by monoclinic orientation and a large optic axial angle ($69-72^\circ$).

All varieties of the potash feldspars can be purely potassic or contain admixtures of the albite molecule up to 40%. This addition can be easily recognized by the increase in refractive index. A diagram correlating refraction and composition is given in Fig. 53.

Potash feldspars (Table 3) are easily decomposed by weathering and by hydrothermal metamorphism resulting in the formation of kaolin (grey or brownish scales with weak double refraction) or sericite.

Plagioclases are isomorphous mixtures of albite $\text{NaAlSi}_3\text{O}_8$ and anorthite $\text{CaAl}_2\text{Si}_2\text{O}_8$.

They are generally distinguished from other feldspars by higher refraction and the presence of polysynthetic twinning (Figs. 54 and 55).

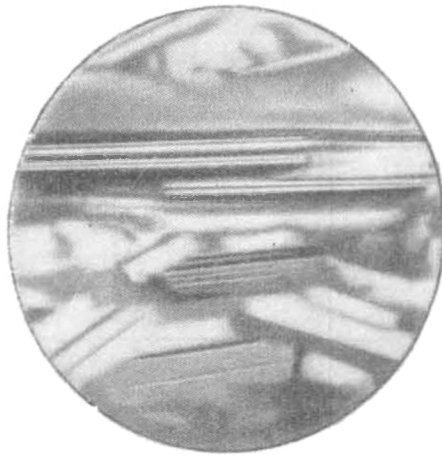


Fig. 55. Plagioclase twins, $\times 21$. Crossed nicols

Table 3

The Main Constants of the Potash Feldspars

Name of feldspar	n_x	n_y	n_z	2V, deg	Twinning	Special features
Microcline	1.525-1.530	1.522-1.526	1.518-1.526	77-84	On the albite and pericline laws	Distinct reticulate structure
Orthoclase	1.526	1.522	1.519	69-72	Karlsbad, Manebach and Baveno	Often perthitic. A brown shade when turbid
Anorthoclase	1.527-1.549	1.526-1.549	1.525-1.532	From 55 to 30-35	On the albite and, more rarely, on the pericline laws	Distinguished from microcline by smaller 2V
Sanidine	1.524-1.526	1.523-1.525	1.517-1.520	0-30	As in orthoclase	Optic axial angle very small (down to 0°)

The most prevalent twin law in the plagioclases is the albite; the composition plane in such twins is the pinacoid (010), while the twin axis is perpendicular to it. One individual of the twin is turned through 180° relatively to the other and the (001) face forms an angle other than 90° with the (010) face. In the same way the faces are inclined to each other as are the axes of the indicatrices of both individuals.

Polysynthetic twinning is obtained as a result of a series of such twins growing together, dark and light individuals being mutually parallel.

E. Fedorov suggested that plagioclases should be designated by numbers (Table 4) corresponding to the anorthite content. For

Table 4

Systematic Nomenclature of Plagioclases

Group	Name of plagioclase	Content, %		No.
		albite	anorthite	
Acid plagioclases	{ Albite	100-90	0-10	0-10
	{ Oligoclase	90-70	10-30	10-30
Intermediate plagioclases	Andesine	70-50	30-50	30-50
	Labradorite	50-30	50-70	50-70
Basic plagioclases	{ Bytownite	30-10	70-90	70-90
	{ Anorthite	10-0	90-100	90-100

example, plagioclase No. 57 signifies a plagioclase containing 57% of anorthite. Along with this, another system of nomenclature exists, in which each mineral of the plagioclase group is denoted by its own name, but to this name a specific meaning is imparted: plagioclases containing less than 10% of anorthite (up to No. 10) are termed albite; from 10 to 30% of anorthite—oligoclase, and so on.

The plagioclases are more conveniently determined by using their increase in refraction index with anorthite content (Fig. 56 and Table 5). For this, the least refractive index (n_z) of the grain studied must be measured by the immersion method. The grain under examination can be taken out of the slice and the Canada balsam washed off with xylol, or a cleavage fragment can be taken from a fresh specimen of the rock. The latter case gives the more accurate results. The measured values are plotted as ordinates; then, using the curve given in Fig. 56, the composition of the plagioclase can be read off along the abscissa.

Table 5

Optical Properties of the Plagioclases

Name	n_x	n_y	n_z	$n_x - n_z$	$2V$, deg
Albite	1.536	1.529	1.525	0.011	+71
Oligoclase	1.5465	1.5428	1.5385	0.0080	-86
Andesine	1.5570	1.5533	1.5500	0.0070	+89
Labradorite	1.5678	1.5629	1.5598	0.0080	+77
Bytownite	1.5760	1.5720	1.5670	0.0090	-83
Anorthite	1.5885	1.5832	1.5755	0.0130	-77

A second, important way of determining the plagioclases is by their optical orientation. The optical indicatrix is oriented differently in the various plagioclases. Fig. 57 gives this orientation projected on to the plane perpendicular to (010), plane (010) itself being given as a vertical line, n_x as a thin line, and n_z as the thicker arrow intersecting the latter. The Fedorov stage is often used for

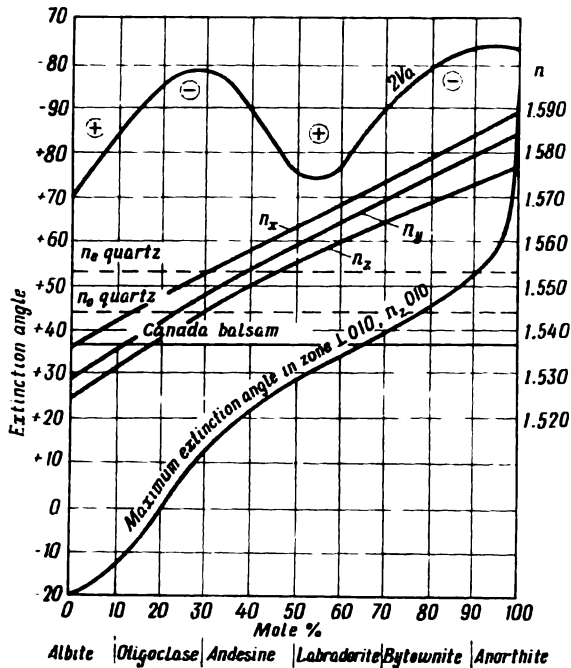


Fig. 56. The optical properties of plagioclases

more accurate determination of plagioclases by finding the position of the optical indicatrix in the crystal.

Many cases can be limited to the determination of the extinction angle, but it is necessary to do this in some strictly defined section. For example, a section is selected perpendicular to both of the best developed pinacoids (010) and (001). These are, of course, the directions of best cleavage, while the (010) pinacoid is, besides, the composition plane of the albite twinning. Therefore, when the (010) plane is arranged perpendicular to the plane of the thin section, the extinction of both individuals is symmetrical, and if this composition plane is placed parallel to either of the cross-hairs in the ocular, then the twinned plate will be uniformly illuminated between crossed nicols and the twinning will seem to have vanished. The perpendicularity of the section to the (001) face is confirmed by the fact that the cleavage cracks are sharp, and do not shift when the microscope tube is raised or lowered (check at high magnification). When such a section has been found, the extinction angle is to be measured between the (010) plane and n_z , and is always less than 45° . The method is complicated by the fact that albite and some other oligoclases give angles between 0° and 16° and very often it is impossible to distinguish between them. It is therefore better to resort to another method.

It is seen from Figure 56, that oligoclase No. 20 possesses an extinction of 0° and that its average refractive index is equal to that of Canada balsam (1.537). Consequently, if the refraction of a plagioclase, having an extinction angle between 0° and 16° , is higher than that of Canada balsam, then this oligoclase must mark its extinction angle with a plus sign; if the refraction is lower, then it must be minus.

The refraction is plotted along the ordinates, and the plagioclase numbers along the abscissa.

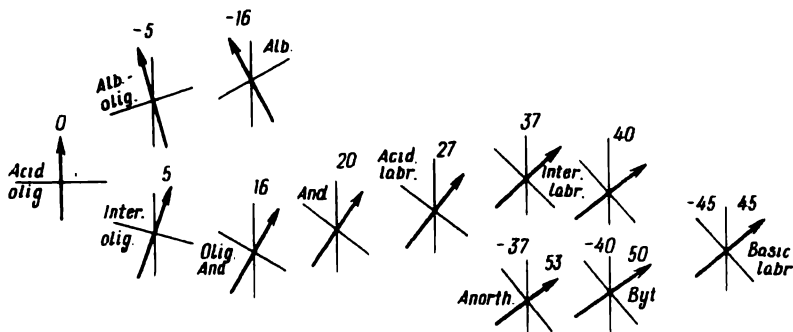


Fig. 57. Maximum extinction angles of plagioclases in sections perpendicular to the composition plane of albite twinning (010)

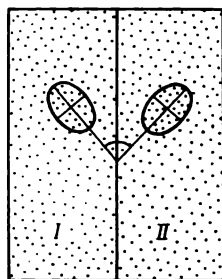


Fig. 58. A section of an albite twin perpendicular to the twin plane

If crystals perpendicular to both (010) and (001) cannot be found, it is possible to limit oneself to section perpendicular to (010). The two inseparable twins (*I* and *II*) in such a section, extinguish at the same angle with the twin suture. On illumination, the tint of each part is barely distinguishable from that of the other, when the twin suture is parallel to a cross-hair (Fig. 58).

The extinction angles are measured in several such sections, and the greatest values are taken into account. The plagioclase number is then determined from the diagram.

It should, however, be noted, that in some cases, different plagioclases can give the same extinction angle; this occurs in the case of acid plagioclases, and when the angles are less than 16° . These cases must then be distinguished as in the previous instance by the refraction and comparing with Canada balsam; in the case of the basic plagioclases, the position of n_z must be checked by the gypsum plate.

Both the preceding methods are inconvenient when studying microlites (microcrystals) of plagioclases in effusive rocks. The length of a microlite is most frequently parallel to the intersection of the two principal cleavages (the edge 100). The extinction angle

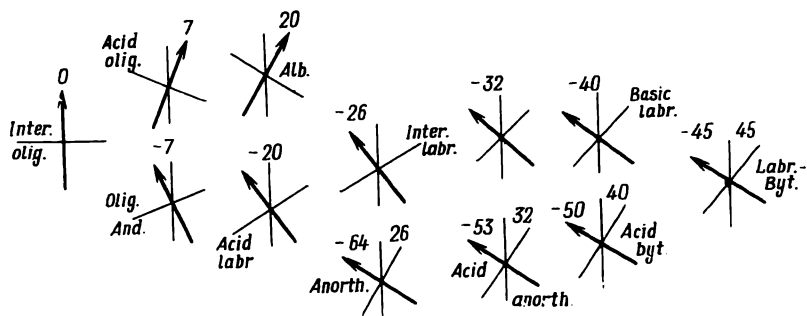


Fig. 59. Maximum extinction angles in longitudinal sections of plagioclase microlites; the vertical line is the trace of the edge (100)

must here be measured relative to the long axis of the microlite, and the greatest angle selected from several (up to ten) microlite measurements. Characteristic maximum angles for several varieties of plagioclase are given on the diagram (Fig. 59). Two points should also be noted in this case: if the extinction angle measured is 26° or greater, the position of n_z must be determined by using the quartz wedge or some other compensator; if the angle is 15° or less, comparison with Canada balsam must be resorted to.

As has already been pointed out, the determination of the plagioclase number is one of the first tasks in the description of a rock. It is also useful to describe the form taken by the crystals and their dimensions.

In some rocks, the crystallization of the plagioclases proceeded in a regular order: the higher melting, anorthite-rich varieties were the first to separate, and then the more acid varieties grew in zones around them, and so on right up to pure albite. Such zoned plagioclases are usually contained in effusive rocks and intrusions which have solidified at no great depth.

Potash feldspars are encountered in rocks rich in alkalis—granites, syenites, liparites.

Plagioclases are an important constituent of practically all magmatic rocks, acid plagioclases in acid rocks, and basic plagioclases in basic rocks. Besides this, albite is often met with as a secondary mineral.

Quartz is one of the most common (12% of the total composition of the earth's crust) and most stable minerals. It is found in effusive and sedimentary as well as in metamorphic rocks. Assuming that its physical properties are well known from mineralogy courses, the data given below will be only what is required for microscopy.

The optical properties of quartz, as well as its composition (SiO_2), are remarkably constant. In thin section, it is usually colourless and transparent. There is no cleavage, and only rarely are dusty inclusions found forming turbid bands. Quartz has no shagreen surface or relief, since $n_e = 1.553$ and $n_o = 1.544$, $n_e - n_o = 0.009$. It is optically positive. Quartz has a higher refractive index than Canada balsam or orthoclase and microcline, and this causes the Becke line at its boundary with these minerals to move into the quartz when the microscope tube is raised. There is a clear dispersion effect at this boundary, the quartz appearing greenish. A thin section of normal thickness of quartz between crossed nicols gives interference colours no higher than white or yellowish-white, while in sections perpendicular to the optic axis (dark in parallel light) in convergent light, it gives a black cross with a yellowish ring on the periphery of the field of view. Its most important properties, its uniaxiality and optic sign, make it easily distinguishable from, for example, albite or nepheline, even in the most obscure cases.

Quartz does not give twins which can be distinguished under the microscope, although quartz twins are found fairly frequently in nature. Usually, the grain extinguishes as a whole at once when the stage is rotated, but sometimes, especially in rocks which have been subjected to pressure, the quartz grains extinguish in waves, or else, at the moment of extinction separate patches of illumination are seen as if it were a mosaic.

Quartz is stable only at temperatures below 870°C. At higher temperatures it passes into tridymite, and at still higher ones, above 1470°C, into cristobalite. But even within the lower temperature range, there is not one, but two quartzes, the low-temperature trigonal, crystallizing as elongated prisms at temperatures below 575°C (all hydrothermal quartz belongs to this type), and the high-temperature, hexagonal, which crystallizes as bipyramidal crystals.

These modifications can be distinguished under the microscope only when the faces are well preserved, but we usually see irregular grains in thin section and the problem can be resolved only from the paragenesis.

Cristobalite and tridymite are found in the cavities of effusive rocks, and, in particular, they are frequently found in silica-rich industrial products which have been subjected to the action of high temperatures, such as acid refractories and acid slags.

Both these minerals are easily distinguished optically by their low refraction and double refraction (for cristobalite, $n_x = 1.484$, $n_z = 1.487$, $n_x - n_z = 0.003$; for tridymite, $n_x = 1.473$, $n_z = 1.469$, $n_x - n_z = 0.004$). A scaly structure is characteristic of cristobalite, and in lavas it sometimes forms spherulites. Tridymite is easily recognized by its low refraction and lanceolate twins.

A very finely fibrous variety of low-temperature quartz, which usually forms spherulites or felt-like aggregates, bears the name of chalcedony. Its structure is well seen under the microscope, and is so characteristic that it is almost impossible to confuse it with other minerals. The refraction and double refraction of chalcedony are usually lower than for quartz ($n = 1.535-1.540$, $n_x - n_z = 0.005 - 0.008$). It is found in lavas where it is an infilling of vesicles in sedimentary rocks (cherts), in hydrothermals, in foundry moulds and flints, and replaces opal in fossil skeletons.

Feldspathoids, or feldspar substitutes. In magmas where there is an amount of alkalis in excess of that required for the formation of feldspars, there may not be enough silicic acid, and then, instead of feldspars, certain minerals rich in alkalis but poor in silica will crystallize, and are known under the name of feldspathoids (Table 6).

The most prevalent of these is nepheline. It forms phenocrysts in lavas as hexagonal prisms, or as rectangular or irregular grains; it is found in syenites and other alkaline rocks. It is very similar to quartz, from which it is distinguished by its negative sign;

Table 6

Feldspathoids

Name	Composition	Hardness	Colour	Crystal system and external aspect	Optical properties	Notes
Leucite	KAlSi_2O_6	5.5	White	Pseudocubic, often isometric grains	Isotropic; $n=1.508-1.509$	Sometimes weakly birefringent
Sodalite	$3\text{NaAlSiO}_4 \cdot \text{NaCl}$	5.5	Blue, pink, white	Cubic, irregular grains	Isotropic; $n=1.48-1.49$	Abundant decomposition products
Hawynne and nosean	Isomorphous mixtures of $6\text{NaAlSiO}_4 \cdot 2\text{NaSO}_4$ and $3\text{CaAl}_2\text{Si}_2\text{O}_8 \cdot 2\text{CaSO}_4$	5.0	Pale blue, colourless	Cubic; cubes, rhombic dodecahedra	Isotropic; $n=1.48-1.50$	Regeneration and inclusions of crystals along the margins are characteristic
Nepheline	NaAlSiO_4 , often up to 20% of KAlSiO_4	5.5	Colourless, pink	Hexagonal prisms, irregular grains	$n_e=1.532-1.544$ $n_o=1.536-1.549$ $n_o-n_e=0.003-0.005$ Uniaxial; negative	—

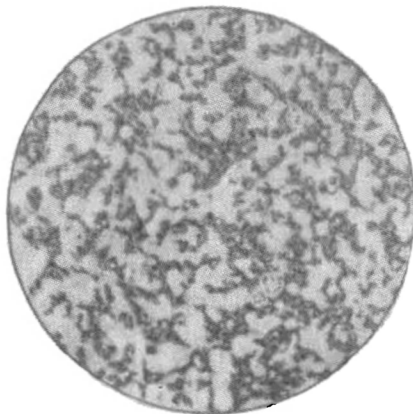


Fig. 60. A leucite rock

cleavage is sometimes seen, and numerous later formed minerals (zeolites, sericite) replacing the nepheline.

Sodalite occurs in conjunction with nepheline, sometimes as its decomposition product.

Hauyne and nosean occur in effusive rocks forming phenocrysts; in thin section, in the form of four-sided or hexagonal sections; cleavage is seldom seen. Their most characteristic features are low refraction and an abundance of inclusions along the edges of the crystals so that it looks like a border.

Leucite is also a mineral of the effusive rocks. It is characterized by a rounded outline (Fig. 60). It is distinguished from analcite which is very similar only by the magnitude of refraction. In old lavas, it is replaced by zeolites or aggregates of nepheline (pseudo-leucite).

Coloured Minerals (Ferromagnesians)

Micas. The most characteristic feature of the micas is an absolutely perfect cleavage in a single direction. All micas give monoclinic (pseudo-hexagonal) crystals, optically negative. The acute bisectrix, which is always n_z , is almost perpendicular to the cleavage lamellae, and sections with clearly seen cleavage possess the greatest birefringence.

Chemically, micas are complex isomorphous mixtures of certain aluminosilicates of aluminium, potassium, lithium or sodium with a sheet structure and always containing much water; continuous sheets of silicon-oxygen tetrahedra and aluminium-oxygen octahedra are bound together by an alkali ion so that the distance between the sheets is far greater than the distance between the ele-

Table 7

Main Differences in the Optical Properties of the Various Micas

Variety of mica	Colour	Pleochroism	n_x	n_z	$n_x - n_z$	$-2V$, deg
Muscovite	Colourless	Non-pleo- chroic	1.58-1.60	1.56-1.58	0.035-0.040	40-70
Phlogopite	Ditto	Ditto	1.565	1.535	0.030	0-10
Biotite	Greenish, brown, black	Strongly pleo- chroic	1.63-1.70	1.58-1.65	0.040-0.060	0-10

ments within the sheet, thus giving rise to the mica cleavage. Part of the aluminium atoms can be replaced by magnesium or iron, the latter either bivalent or trivalent. Micas with sodium or lithium are rather rare in nature, so that only the potash micas are of any importance as rock-forming minerals. Micas containing aluminium bear the name of muscovites, the magnesian micas—phlogopites, and the ferriferous—lepidomelanes.

The magnesian and iron varieties mix in all proportions, forming the various micas known as biotite. Muscovites have a more or less constant composition, and it is only rarely that a small amount of the iron component is present.

The muscovite and phlogopite components apparently do not mix.

The optical properties, like the chemistry, are more or less constant in muscovite and vary a great deal in biotite. In thin section, muscovite is colourless, biotite is coloured the more strongly the greater its iron content. The most important differences are set out in Table 7.

The small optic axial angle in biotite and phlogopite results in their cleavage flakes not having interference tints (almost isotropic), and in convergent light the conoscopic picture resembles that of a uniaxial crystal; the same sections of muscovite, however, have sufficient birefringence for thin sections of normal thickness to give a yellow tint between crossed nicols.

In convergent light, they usually give a good picture of a section perpendicular to the acute bisectrix, and the emergence of both optic axes.

Muscovite, $KAl_2(AlSi_3O_{10})(OH)_2$, is found in metamorphic rocks, especially in mica schists and gneisses, more rarely in granites and pegmatites. Secondary formations of muscovite often occur, for example, replacing feldspars. On weathering it is converted to kaolinite. Typical of muscovite are the perfect single cleavage, the weak relief and shagreen surface, straight extinction.

Between crossed nicols the interference colours are pure second and third order tints—sharp blues, greens and reds.

A fine scaly muscovite is called *sericite*.

Biotite occurs as pseudohexagonal lamellae and little columns. It is found in granites, mica schists, and alters to chlorite; it sometimes replaces amphiboles and pyroxenes.

The pleochroism (from light yellow to dark brown) is the most characteristic feature of biotites, even of the pale coloured ones. The darkest colour is seen when the polarizer vibration direction is parallel to the cleavage or to the long direction of the section.

Amphiboles and pyroxenes. Minerals of the general formula $RSiO_3$, where $R = Mg, Ca, Fe^{II}, Fe^{III}, Al$. The presence of water or the volatiles (F, Cl) is obligatory in amphiboles, but they are absent in pyroxenes.

Both groups of minerals are characterized by elongated prismatic crystals with two good cleavages along the prism (110). In slice, sections parallel to the vertical axis can be found showing one direction of cleavage, and sections perpendicular to this axis show the intersecting cleavage.

There are orthorhombic and monoclinic members of both mineral groups, the monoclinic amphiboles and pyroxenes being the most widespread. Orthorhombic amphiboles are very rare; orthorhombic pyroxenes are of fairly frequent occurrence. The orthorhombic varieties are distinguished from the monoclinic by straight extinction in all the principal sections. The differences between amphiboles and pyroxenes are shown below:

	<i>Amphiboles</i>	<i>Pyroxenes</i>
Angle of prismatic cleavage, deg	About 124 (see Fig. 2)	About 87 (see Fig. 21)
Colour	Dark	Colourless
Pleochroism	Strongly pleochroic	Non-pleochroic
Birefringence	Small	Bright interference colours
Extinction angle of n_x	Small ($12-25^\circ$) with rare exceptions	Large ($30-45^\circ$)
Optic axial angle	Large	Medium
Sign	Negative	Positive
Alteration	Is converted into chlorite, biotite, etc.	Is converted into amphiboles, chlorite, biotite, etc.

The most important of all these features are the angle made by the prismatic cleavage and the extinction angle. If there is any doubt as to whether a mineral belongs to the pyroxenes or amphi-

boles, a section should be found showing the intersecting cleavages; an angle around 120° indicates an amphibole, about 90° , a pyroxene.

The extinction angles should be measured in sections perpendicular to the optic normal; these are easily recognized by the parallel cleavages and by having the highest interference tint.

Certain additional data should also be taken into consideration.

The pale green monoclinic amphiboles, containing Mg and poor in Fe, form the actinolite series.

The darker ones with higher Fe content and containing Ca form the hornblende series; those especially rich in Fe and containing alkalis are called riebeckites.

Refractive indices: for actinolite $n_x = 1.659$, $n_z = 1.638$; for common hornblende $n_x = 1.65$ - 1.68 , $n_z = 1.63$ - 1.66 ; for riebeckite $n_x = 1.697$, $n_z = 1.693$.

The distinct prismatic cleavage at an angle about 124° is characteristic.

The birefringence is 0.023 and increases with iron content. The bright interference colours are somewhat distorted by the mineral's own tint.

Twins occur, and are sometimes polysynthetic. Pleochroic haloes are seen around inclusions.

Amphiboles enter into the composition of syenites, diorites and amphibolites. They are easily decomposed, but often turn up in a fresh condition in sedimentary rocks, mainly as sands and sandstones.

For pyroxenes, a cleavage at an angle of 87 - 88° is characteristic, but it is less perfect than in the amphiboles. In the orthorhombic and some of the monoclinic pyroxenes, a pinacoidal (along 100) parting is added to the prismatic cleavage, as a result of which they assume a foliate resemble (diallage). In the piece, they are usually coloured (green, brown, black). Short columnar crystals are frequent.

The usual pyroxenes, the augites, are the commonest dark minerals of the igneous rocks, especially pyroxenites, peridotites, diabases, gabbros, monzonites. The alkaline pyroxenes—*aegirines* are often encountered in nepheline syenites and in their pegmatites.

Pyroxenes are often converted into chlorite and hornblende.

The most common monoclinic pyroxenes are those of the diopside-augite series comprising: diopside $\text{CaMg}(\text{SiO}_3)_2$, hedenbergite $\text{CaFe}(\text{SiO}_3)_2$, augite—an isomorphous mixture of diopside, hedenbergite and MgSiO_3 with a significant admixture of the corresponding isomorphous aluminous and titaniferous molecules.

The optical properties of these minerals vary with the iron content and the ratio Mg—Ca approximately as shown in Fig. 61. But even when the optical properties of a pyroxene have been determined, it is difficult to say anything about its composition, because

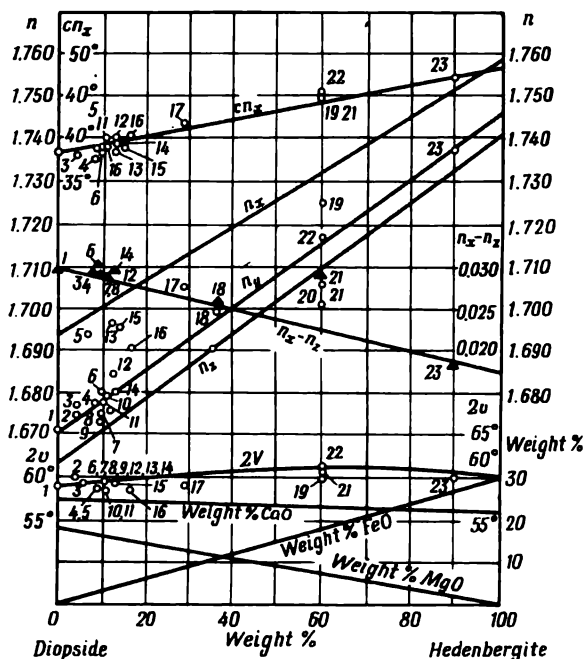


Fig. 61. Optical properties of monoclinic pyroxenes

there is always some effect due to other additions not shown on the diagram.

The birefringence of augite is fairly strong ($n_x - n_z = 0.020 - 0.030$); the extinction angle is $38-54^\circ$; twinning is frequent; augite is biaxial, positive. The emergence of one optic axis is usually seen in sections perpendicular to a prism. The alkaline pyroxene, aegirine, is somewhat different from augite, being characterized by the pyroxene cleavage angle, high refraction $n_x = 1.836$, $n_z = 1.776$, but differing from other pyroxenes by its bright green colour and its pleochroism in tones of yellow and green. The extinction angles cn_z are small, up to 7° (cn_x is very large).

The optic sign is negative.

The orthorhombic pyroxenes have more definite properties, since they form a mixture of only two original components: enstatite MgSiO_3 and ferrosilite FeSiO_3 .

The latter is rarely found in nature but is formed in some slags.

The optical properties vary with increasing content of FeSiO_3 , as shown in Fig. 62, and so can be used (especially 2V) to determine the composition of an orthorhombic pyroxene.

All orthorhombic pyroxenes are characterized by straight extinction and prismatic cleavage at an angle about 88° .

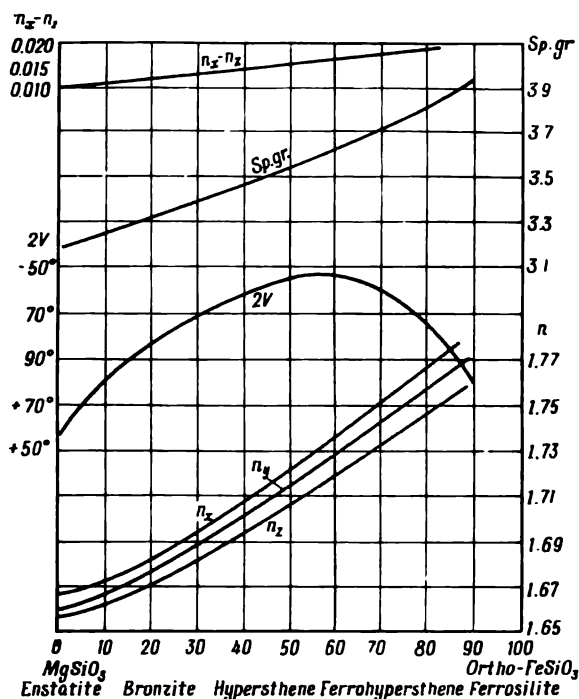


Fig. 62. Optical properties of orthorhombic pyroxenes

Enstatite and bronzite are colourless, hypersthene is feebly coloured and pleochroic in shades of green and brown.

The relief is fairly high; for enstatite $n_x = 1.665\text{--}1.671$, $n_z = 1.656\text{--}1.661$, for hypersthene $n_x = 1.700\text{--}1.731$, $n_z = 1.692\text{--}1.715$.

The birefringence of enstatite is weak (0.009); average for bronzite and hypersthene (0.013); but can rise to 0.020. The elongation is always positive and is very characteristic.

Enstatite is positive, with a large optic axial angle; hypersthene and bronzite are negative. The emergence of a positive bisectrix can be obtained in convergent light in sections perpendicular to the prism, in distinction to monoclinic pyroxenes of the diopside series.

The olivine group. A large group of rock-forming minerals of general formula $R_2\text{SiO}_4$, where R is mainly Mg and Fe²⁺, but Mn²⁺ and other bivalent elements can be present in insignificant amounts. They crystallize in the orthorhombic system forming characteristic crystals which are combinations of bipyramids and pinacoids, and very typical grains, the product of the melting of the crystals. Often replaced by serpentine along cracks. Transparent or translucent, the colour in the specimen is bottle-green or black.

The following varieties are distinguished according to Fe_2SiO_4 content:

	Fe_2SiO_4 , %
Forsterite	0-10
Olivine	10-30
Hyalosiderite	30-50
Hortonolite	50-70
Ferrohortonolite	70-90
Fayalite	90-100

The most common are the olivines themselves, being found both in gabbros, peridotites and many other intrusive rocks, and in effusives (basalts, andesites). Fayalite is a habitual component of slags in non-ferrous metallurgy.

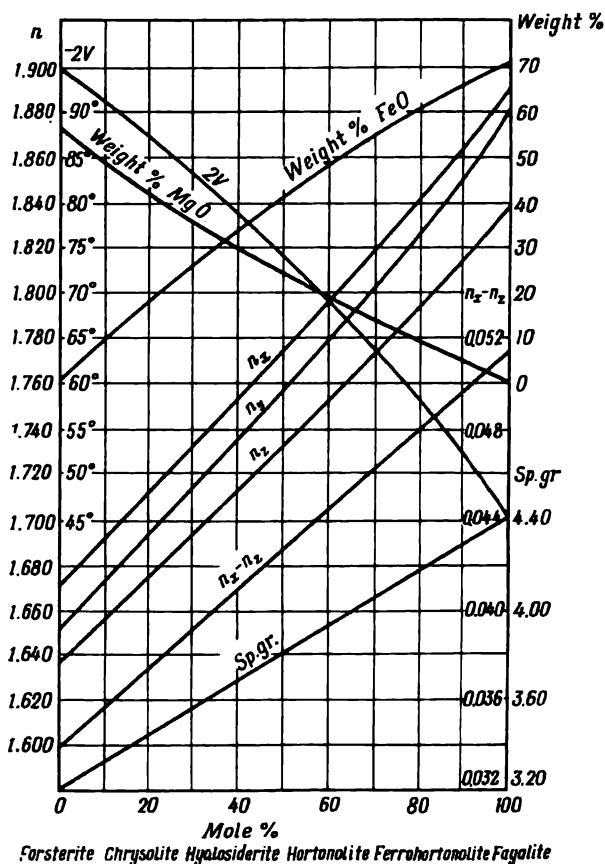


Fig. 63. Optical properties of olivine

On alteration, olivine is converted to serpentine or to minerals like it.

The optical properties of the olivines vary with the content of ferrous molecules. Measurements of the refraction or of $2V$ enable the composition of the mineral to be determined from the data given in Fig. 63.

The strong relief, pitted surface and high refractive index ($n_x = 1.686-1.709$; $n_z = 1.651-1.672$) are typical of olivine. Colourless in thin section. Cleavage is usually absent, but is sometimes excellent in basalts.

Between crossed nicols there is bright interference colouring due to the strong birefringence (0.036-0.050). The extinction is straight, observable when the external form of the crystal is visible. Biaxial, negative. The angle $2V$ is large; a black band almost without curvature in convergent light.

Olivine is distinguished from pyroxenes and amphiboles by the absence of a distinct cleavage, by the generally higher refraction, as well as its alteration to serpentine along cracks. Serpentine is characterized by weak birefringence and fibrous structure, the fibres being arranged perpendicular to the cracks.

2. THE LESS IMPORTANT CONSTITUENTS OF MAGMATIC ROCKS

Accessory Rock-Forming Minerals

These minerals are often encountered in rocks, but as rare grains. In most cases, their amount does not exceed 0.5-1%, and it is only in very rare instances that they are plentiful.

The accessory minerals are conveniently arranged according to their optical properties: first the opaque ones, then the isotropic and birefringent ones, in the order of increasing birefringence.

Opaque minerals. Magnetite, titanomagnetite, chromite, pyrite and other ore minerals belong to this group.

Magnetite ($\text{FeO} \cdot \text{Fe}_2\text{O}_3$). Cubic system. Iron-black in colour with metallic lustre. Magnetic. Hardness 5.5. It is often found in the most diverse rocks. Opaque under the microscope. Forms octahedra or irregular grains. In reflected light (if the section is lit from above, the light source is first raised) blue-black with a distinct metallic lustre.

Titanomagnetite. Magnetite with some titanium. Under the microscope it can be distinguished from magnetite only by that titanite grains are formed around a grain of titanomagnetite on decomposition as a fine-grained aggregate (leucoxene). It is found in basic rocks.

Chromite ($\text{FeO} \cdot \text{Cr}_2\text{O}_3$). Iron can be replaced by magnesium and chromium by aluminium, resulting in the formation of a large group known as the chrome-spinellids. Distinguished from magnetite by becoming slightly translucent in strong light. Found in ultrabasic rocks.

Pyrite (FeS_2). Cubic system. Hardness 6.5. Cubes, pentagonal dodecahedra. No cleavage. Colour in specimen is straw-yellow. Metallic lustre. Opaque under the microscope. In reflected light—yellow (which distinguishes it from magnetite and other opaque minerals). Found in the most diverse rocks.

Transparent minerals. Fluorite, apatite, tourmaline, zircon, sphene belong to this group.

Fluorite (CaF_2). Cubic system. Often cubes. Cleavage along the octahedral faces. Hardness 4. Found in igneous rocks (mainly alkaline) and hydrothermal veins, sometimes forms the cement in sandstones. Under the microscope is distinguished by two or three systems of cleavage cracks and by being isotropic. Very low refractive index, 1.43, and distinct negative relief. Colourless or bluish or violet in colour.

Apatite [$\text{Ca}_5(\text{F}, \text{Cl})(\text{PO}_4)_3$]. Constantly found in igneous and frequently in metamorphic rocks, more rarely in sedimentary rocks. The main constituent of phosphorite.

Under the microscope, apatite most frequently appears as little hexagonal prisms, and as isotropic hexagons in cross section. It is usually colourless, but is sometimes coloured and is then pleochroic.

The refractive index $n_o = 1.632$ – 1.648 , $n_e = 1.630$ – 1.643 causes strong relief. Weakly birefringent: 0.002 – 0.005 —a grey colour of the first order. Characteristically uniaxial and with negative elongation.

Tourmaline. A boro-alumino-silicate containing Na, Li, Mg and Fe. Hexagonal (trigonal). Columnar crystals striated along their length; transverse sections triangular. Almost colourless, or more often coloured blue, pink or green. Zoning is sometimes seen, the one grain showing different colours, and a sudden change in colour around inclusions.

Coloured grains are pleochroic, the tourmaline being faintly coloured when the vibration direction of the polarizer coincides with n_e , the elongation direction of the mineral.

The refractive index is about 1.63, giving strong relief, and the strong birefringence, 0.02 , gives bright interference colours. Straight extinction. Uniaxial, negative.

Found in acid rocks, crystalline schists, and when these disintegrate, in the resulting sands and clays.

Zircon (ZrSiO_4). Tetragonal. Small, well-formed crystals (Fig. 64). Colour from pale yellow to dark brown. Forms stumpy prisms in rocks; elongated prisms are rare; quadrangular or octagonal in cross section. Colourless in thin section. Extraordinarily high

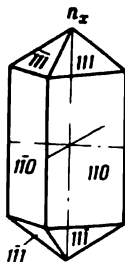


Fig. 64. Tetragonal prism and bipyramid of zircon

refraction, $n_o = 1.923-1.960$; $n_e = 1.968-2.015$, very high relief, birefringence from 0.044 to 0.062 gives bright interference colours.

If a zircon is enclosed in a coloured mineral, a darker colour is often observed directly around the zircon inclusion, forming a pleochroic halo as a result of its radioactivity (UO_2 or ThO_2 replaces ZrO_2 isomorphously).

Zircon is easily distinguished from apatite by high interference tint.

It is found in granites, syenites, gneisses, mica schists and in sands and sandstones resulting from their decomposition, but without suffering any decomposition itself.

Sphene (titanite) ($CaTiSiO_5$). Monoclinic. Hardness 5-5.5. Flat, wedge shaped crystals, often of a singular habit, reminiscent of an envelope, more rarely prismatic. Under the microscope greyish, more rarely of several shades of brown. Sometimes weakly pleochroic. Characteristic features are the crystal form and very high relief resulting from the high refraction: $n_x = 2.034$, $n_z = 1.900$.

The exceptional value of the birefringence (0.121 and higher) gives rise to white interference colours of the highest order, which distinguish it from zircon. Oblique extinction. Biaxial, positive, with small $2V$.

Found in most intrusives and in many metamorphic rocks.

Secondary Minerals

All the minerals listed in this section are products of the decomposition of various primary minerals or are formed by crystallization from hydrothermal solutions. Some of them are essential constituents of metamorphic (epidote, chlorite) or sedimentary (calcite, opal) rocks. The list has been compiled in accordance with their occurrence in nature, with the exception of calcite, which is described below as a typical mineral of sedimentary rocks.

The epidote group. One of the minerals of this group, *zoisite*, has the composition $3\text{CaAl}_2\text{Si}_2\text{O}_8 \cdot \text{Ca}(\text{OH})_2$, in which Al can be replaced by iron, and in part by rare earths (orthite is an accessory mineral of the acid rocks). Zoisite usually has an orthorhombic form, but also exists as the monoclinic variety, clinozoisite. If the mineral contains more than 5% of the ferriiferous molecules, then the only stable form is the monoclinic. These iron rich varieties bear the name of *epidote*. The properties of the monoclinic varieties vary systematically with the content of the ferriiferous molecules, as shown in Fig. 65.

More detailed data for typical varieties are given below.

Zoisite. Orthorhombic. Usually found as an alteration product of plagioclase in the form of short prisms or irregular grains.

Typical mineral of green schists. Under the microscope, the high refractive index (1.703) causes sharp relief. Colourless in thin section. Cleavage is sometimes seen in the form of some parallel cracks. Weak birefringence (0.005), and anomalous interference colours resulting from the strong dispersion of the birefringence.

The conjunction of very weak birefringence with high refractive

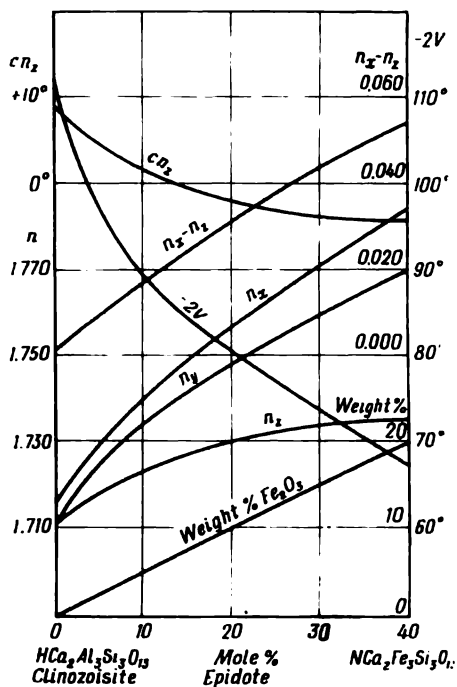


Fig. 65. Optical properties of epidotes

index and anomalous blue or reddish interference tints makes zoisite easily distinguishable from all other minerals.

Epidote. Monoclinic. Coarsely striated, prismatic crystals as in tourmaline. Perfect cleavage. Colour is pistachio-green to dark green, almost black.

Microscopic grains of epidote in rocks give them a green colour.

Under the microscope colourless or yellowish, pistachio-green. The brightly coloured varieties are pleochroic, from colourless to yellow (very characteristic of epidote). Angle between cleavages is 65° (in distinction to the monoclinic pyroxenes). The refraction is very high, $n_x = 1.768$, $n_z = 1.729$, causing strong relief. Birefringence, 0.037, which in conjunction with the absorption is the cause of the characteristic particoloured (a distinction from pyroxene) interference colours (yellow-orange, bright crimson, greenish-blue). Biaxial, negative.

Epidote often replaces decomposed plagioclases in basic magmatic rocks; it is also found in green and chlorite schists, in phyllites, amphibolites and sandstones.

Chlorite. Hydrated silicate of Mg, Fe and Al. Green colour. Similar to mica, but the lamellae are not elastic. Monoclinic, pseudohexagonal.

The optical properties vary with the iron content, but are always a diagnostic feature. Cleavage perfect in the highest degree. Straight, or almost straight extinction relative to the cleavage.

The colour and pleochroism is in shades of green; weak birefringence (0.001-0.011) and anomalous interference tint (dark blue, violet and lilac colours as a consequence of the strong dispersion). Refraction: $n_x = 1.577$ -1.610; $n_z = 1.562$ -1.605.

The anomalous tint and low birefringence easily distinguish chlorite from the green biotites. Often found as a secondary mineral.

Found in igneous rocks, crystalline schists, in contact rocks often associated with ore deposits, as an alteration product of biotite, pyroxenes, amphiboles, volcanic glass and other ferromagnesian minerals in "greenstones" (diabases, porphyrites, etc.). In metamorphic rocks, it sometimes constitutes entire beds (chlorite schists).

Serpentine $[H_4(Mg,Fe)_3Si_2O_9]$. Monoclinic. Green; frequent inclusions of black chromite. Dense masses with splintery fracture. Formed from the alteration of olivine and augite rocks. The fibrous varieties are called chrysotile, the platy, antigorite.

Under the microscope, serpentine is colourless or greenish, pleochroism is weak or not noticeable (in distinction to the chlorites). The refractive index varies from 1.490 to 1.551. Birefringence is 0.002-0.011, giving a bluish-grey interference colours.

Zeolites. A large group of aluminosilicates of Ca and Na. The majority are biaxial. Colour white. Analcime and natrolite are the commonest.

Analcime ($\text{Na}_2\text{Al}_2\text{Si}_4\text{O}_{12} \cdot 2\text{H}_2\text{O}$). Cubic. The form is reminiscent of leucite, and it is referred to this group by most investigators. Forms compact masses, sometimes with cubic cleavage.

Natrolite ($\text{Na}_2\text{Al}_2\text{Si}_3\text{O}_{10} \cdot 2\text{H}_2\text{O}$). Orthorhombic. Forms radiating intergrowths of acicular crystals.

Zeolites replace feldspars and feldspathoids; they are often found in the vacuoles of lavas, especially in basalts.

Under the microscope, zeolites are colourless in the majority of cases, fibrous, acicular, sometimes in radiating masses or as sheaf-like scaly aggregates. Analcime forms compact masses, sometimes with cubic cleavage.

The refractive index is always lower than that of Canada balsam, and sometimes very low, giving a noticeably negative relief. The birefringence is mostly weak (0.003-0.007); analcime is isotropic or with anomalous weak birefringence.

3. MINERALS OF THE SEDIMENTARY ROCKS

Practically all the minerals known in magmatic rocks can be found in sedimentary rocks, occurring there as sediments accumulated from transported fragmentary material; however, along with them are found certain specific newly formed minerals, the most common of which are listed below.

Carbonates. A large group of minerals of general formula RCO_3 , where R can be Ca, Mg, Fe^{++} , more rarely Zn, Co, and so on. A group apart, but very similar, are double salts of the type $R^1R(\text{CO}_3)_2$, where R^1 is Ca, and R , Mg or Fe. They all crystallize in the hexagonal (trigonal) system and possess absolutely perfect rhombohedral cleavage with similar angles. The extraordinarily high birefringence is the most characteristic of the optical properties.

The rock-forming members of this group are calcite and dolomite.

Calcite (CaCO_3). Hexagonal. Colourless in thin section. Forms irregular, sometimes palmate grains. Perfect rhombohedral cleavage and polysynthetic twinning are well seen.

The refractive index $n_o = 1.658$, $n_e = 1.486$, and the unusually high birefringence (0.172) give rise to what may be called pseudoabsorption in the presence of one nicol only, i.e., the mineral is almost colourless in one direction (along n_e) and has a smooth surface; in the other (along n_o), it has a shagreen surface and is greyish; between crossed nicols, there are very delicate iridescent interference colours. Uniaxial, negative, straight extinction (along the diagonal of the rhomb). Coloured rings are seen in convergent light, even in a slice of normal thickness.

It is the sole constituent of certain marbles and limestones. When mixed with clay forms marls. Often occurs as the cement in

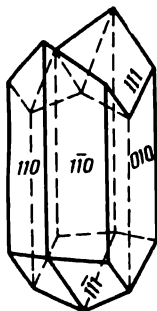


Fig. 66. Swallow-tail twin of gypsum

sandstones, sometimes as a decomposition product of other minerals.

Dolomite is a double salt ($\text{CaCO}_3 \cdot \text{MgCO}_3$). Hexagonal. As distinct from calcite, it does not effervesce with HCl at normal temperature. Very similar to calcite, frequently forming well developed rhombohedra. Twins are rare. Refractive indices $n_o = 1.679$, $n_e = 1.502$, therefore, just as in calcite, pseudoabsorption can be seen: in one direction along n_e the crystal is almost colourless with a smooth surface, while in the other (along n_o) it has the shagreen surface and is greyish.

It is difficult to distinguish dolomite from calcite; in thin sections well developed faces, the absence of twins and the higher refraction are characteristic. In doubtful cases, it is better to resort to polished sections and reflected light; after etching with weak acid in the cold, it is immediately obvious that dolomite is not etched. It often forms individual beds or layers, is sometimes found in limestones and gypsum.

Kaolinite ($\text{Al}_2\text{O}_3 \cdot 2\text{SiO}_2 \cdot 2\text{H}_2\text{O}$). Monoclinic, pseudo-hexagonal. Scaly segregations or platy crystals are seen under the microscope, also vermicular intergrowths. Perfect cleavage, like that of mica. Colourless, turbid. Refractive indices $n_x = 1.566$, $n_z = 1.560$. Birefringence 0.006-0.007. Grey interference tint (in distinction to muscovite).

Gypsum ($\text{CaSO}_4 \cdot 2\text{H}_2\text{O}$). Monoclinic. Twins are frequent (Fig. 66). Fibrous gypsum, selenite, is used as a cheap adulterant. Under the microscope, colourless, similar to quartz, but frequently coloured by iron oxides. Low refractive indices, $n_x = 1.529$, $n_z = 1.520$. Low birefringence, 0.009. In distinction to quartz, biaxial; optically positive. Cleavage is sometimes seen, and then the extinction is equal to $13-37^\circ$.

Glaucinite. A hydrated aluminosilicate of K and Fe, near biotite. Granular, more often earthy. Colour green, brown on weathering. Monoclinic. The refractive index depends on the iron content and

varies in a wide range; mostly around 1.61. The shagreen surface is clearly seen. Extinction is almost straight. Spheroidal aggregates. In foliate varieties is sometimes pleochroic. Rather bright interference colours between crossed nicols. Found exclusively as a new mineral in sedimentary rocks, sandstones and limestones.

Opal ($\text{SiO}_2 \cdot n\text{H}_2\text{O}$). Amorphous. Under the microscope, it is characterized by being colourless, by the absence of cleavage and by irregular cracks. The refractive index is very low (1.40-1.46). This causes a golden-yellow dispersion effect at its boundary with balsam, and low relief. Isotropic; sometimes anomalous grey interference colours. Usually incrustations and compact masses. Found in hot-spring deposits. Forms the cement of sandstones and diatom shells, and also occurs as rounded bodies in tripoli earths.

Limonite ($\text{Fe}_2\text{O}_3 \cdot n\text{H}_2\text{O}$). Colour from ochre-brown to black. Usually opaque, translucent in very thin grains and yellow. Yellow in reflected light; found as incrustations and spongy masses when a decomposition product of iron minerals, and as the cement of sandstones, and also forms independent beds (for example, bog iron).

4. MINERALS OF THE METAMORPHIC ROCKS

The main mass of minerals forming metamorphic rocks comprise minerals such as feldspars, micas, chlorite, and, very commonly, the amphiboles as well as others which have already been described, but besides these, some minerals are known to occur almost exclusively in metamorphic rocks. As in the case of the minerals just enumerated, the most important and typical minerals only of this new group will be dealt with below.

The garnet group. A large group of silicates belonging to the cubic system with the general formula $R_3^+R_2^{++}(\text{SiO}_4)_3$, where R_3^+ is Mg, Fe, Mn, Ca; R_2^{++} is Al, Fe, Cr.

The following basic types of garnets are distinguished:

Almandine	$\text{Fe}_3\text{Al}_2(\text{SiO}_4)_3$	Red
Pyrope	$\text{Mg}_3\text{Al}_2(\text{SiO}_4)_3$	Colourless, red
Spessartine	$\text{Mn}_3\text{Al}_2(\text{SiO}_4)_3$	Brown, black
Grossular	$\text{Ca}_3\text{Al}_2(\text{SiO}_4)_3$	Colourless, or brown
Andradite	$\text{Ca}_3\text{Fe}_2(\text{SiO}_4)_3$	Black
Uvarovite	$\text{Ca}_3\text{Cr}_2(\text{SiO}_4)_3$	Bright green

The pure form of any type of garnet is not found; each actual garnet usually represents a mixture of two or more components, with one or two components distinctly dominant, and the remainder occurring in small amounts (less than 5% of each). The optical

properties of garnets also vary, within narrow limits, with change of composition.

All garnets possess a high refractive index (1.76-1.83), and consequently have a shagreen surface and positive relief (thick black outlines). In thin section mostly pinkish and brownish shades; garnets are rarely colourless. No cleavage is seen.

Usually isotropic. In anomalous garnets, grey interference colours and polysynthetic twins intersecting at various angles can be seen.

Almandine is a habitual component of crystalline schists and gneisses. Pyrope is found in metamorphic serpentine rocks. Spessartine is an accessory mineral in some magmatic rocks and metamorphosed manganese ores. Uvarovite is a component of chromium ores. Grossular and andradite, which usually form isomorphous mixtures in all proportions, are found in contact-metamorphosed limestones (skarns).

Talc $[\text{Mg}_3(\text{OH})_2\text{SiO}_4\text{O}_{10}]$. Monoclinic. Talc usually forms plates and scales possessing a mica-like cleavage. In thin section, it is very similar to muscovite; colourless or slightly brown. Its refractive indices vary: $n_x = 1.575-1.590$, $n_z = 1.538-1.545$. The birefringence can reach 0.050 (greater than muscovite). Optically negative.

Under the microscope, it is hard to distinguish from muscovite, it being necessary to make an accurate determination of the constants or a chemical analysis.

Found in metamorphic rocks as a constituent of talc and talc-chlorite schists.

Kyanite (disthen) $(\text{Al}_2\text{SiO}_5)$. Triclinic. In rocks forms columnar or prismatic grains and crystals. Perfect cleavage along the first pinacoid (100); distinct along the second pinacoid (010); parting along the third pinacoid. Usually colourless in thin section; sometimes slightly bluish (a very important feature, as this shade of pale blue is not found in other minerals). High relief, $n_x = 1.728$, $n_z = 1.712$; the birefringence 0.012-0.016, giving yellow interference colours. Optically negative. Elongation is usually positive. Polysynthetic twins are often seen. A mineral characteristic of crystalline schists.

Sillimanite $(\text{Al}_2\text{SiO}_5)$. Usually contains some Fe_2O_3 . Orthorhombic. Fine acicular crystals and fibrous masses, often elongated prisms and fine hair-like fibres; rhombs (almost squares) in cross section. Colourless. The refringence is high: $n_x = 1.68$; $n_z = 1.66$, giving high relief. Straight extinction; optically positive; elongation positive (as distinct from andalusite). Birefringence $n_x - n_z = 0.022$, giving high interference tints. The crystals are sometimes so small that they do not occupy the whole thickness of the slice and, consequently, do not give a large path difference and bright colours. A typical contact mineral, more rarely found in crystalline schists.

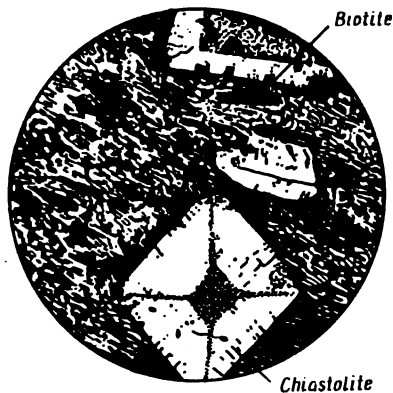


Fig. 67. Chialstolite (andalusite) slate, $\times 20$. Typical arrangement of andalusite inclusions can be seen. Another mineral seen is biotite

Andalusite (Al_2SiO_5). Prismatic crystals of orthorhombic symmetry. Sometimes contains inclusions of carbonaceous materials, which form a cross in transverse section, and in longitudinal section, a certain carbonaceous band (chialstolite, Fig. 67).

The crystals occur as irregular grains and also prisms giving parallelograms and rhombs in cross section. Weak prismatic cleavage at 88° ; colourless or pale pink and pleochroic (very characteristic). Any colour present is distributed irregularly. Refractive indices: $n_z = 1.632$; $n_x = 1.643$, causing noticeable relief. Birefringence almost that of quartz ($n_x - n_z = 0.011$). In sections of normal thickness the interference tint is no higher than the yellow of the first order. If attenuated grains are formed, the elongation is nearly always negative and the extinction straight. Optically negative. Found only in contact metamorphic rocks.

Staurolite $[\text{Fe}(\text{OH})_2 \cdot 2\text{Al}_2\text{SiO}_5]$. Orthorhombic. Prismatic crystals, cruciform intergrowths. Under the microscope, cleavage is distinct; characteristic colour and pleochroism from orange or golden-brown to colourless. The refringence varies, most often $n_x = 1.746$, $n_z = 1.736$, giving considerable relief and a shagreen surface. Birefringence 0.01, consequently, the interference is the same as in quartz. Cruciform twins at a right angle or at an angle of about 60° are characteristic; straight extinction and positive elongation. Optically positive. Large $2V$. A mineral characteristic of the crystal-line schists.

Cordierite $[(\text{MgFe})_2\text{Al}_4\text{Si}_5\text{O}_{18}]$. Orthorhombic. Refractive index varies in the range $n_x = 1.539-1.570$; $n_z = 1.532-1.552$. Birefringence near that of quartz 0.009-0.010 and only rarely is as low as 0.003. Cleavage is only rarely seen, usually colourless in

slice. Polysynthetic twinning in the form of sector trillings and sixlings is characteristic, and also the lemon-yellow pleochroic haloes surrounding radioactive inclusions. Biaxial, optically negative. Under the microscope difficult to distinguish from quartz; the presence of the haloes helps, as also the twinning and the decomposition products (chlorite or iron minerals near it) which never occur in quartz. Should doubt arise, a check whether the mineral is uniaxial or biaxial will definitely decide where the mineral belongs.

Found in contact metamorphic rocks. Usually separates as irregular grains.

5. THE MINERALS OF TECHNOLOGY

Mention has been made earlier of certain very important minerals formed in industrial processes, such as cristobalite and tridymite which are found in acid refractories and acid slags, and fayalite, the iron member of the olivine group, typical of slags in non-ferrous metallurgy.

Here we shall dwell on *monticellite*, which is also characteristic of slags; on *mullite*, a porcelain mineral, very near sillimanite, and on *alite* and *belite*, minerals in cement clinker.

Monticellite (CaMgSiO_4). Orthorhombic. Irregular grains or prismatic crystals. Colourless. Distinct shagreen surface. Close to olivine, from which it is distinguished by low interference tint. Optically negative, $2V = 75^\circ$, $n_x = 1.66$, $n_z = 1.67$.

Found in nature at the contact of limestones with traps. Artificial monticellite, rich in magnesium, is found in some blast furnace slags, and the iron manganese monticellite isomorphous with it, $(\text{Fe}, \text{Mn})_2\text{SiO}_4 \cdot \text{Ca}_2\text{SiO}_4$, in martensite slags and sometimes in dinas.

Mullite ($3\text{Al}_2\text{O}_3 \cdot 2\text{SiO}_2$). Orthorhombic. Colourless. Crystallizes in little prisms with the angle $(110) : (110) = 89^\circ$ and in hair-like needles. Has a distinct cleavage along (010) . The refringence $n_x = 1.654$, $n_z = 1.642$ gives sharp relief. Optically positive, $2V = 45\text{--}50^\circ$. An essential constituent of grog (chamotte) and porcelain. It is also formed when kyanite, andalusite or sillimanite is heated to a high temperature.

Alite ($3\text{CaO} \cdot \text{SiO}_2$). The main constituent of cement clinker, determining its binding properties. In slice, it is colourless and weakly polarizing. Refringence $n_{uv} = 1.720$. Uniaxial, optically positive.

Belite (β . $2\text{CaO} \cdot \text{SiO}_2$). Pale yellow, brightly polarizing grains. Gives polysynthetic twins. Refringence $n_x = 1.736$, $n_z = 1.717$. Biaxial, optically negative. Crystallizes in portland cement clinker. Also occurs in very calcareous blast furnace slags, where on cooling

Table for the Approximate Determination of Some of

		Colourless (sometimes grey. turbid)	
	Amorphous (crypto-crystalline)	<i>Opal</i> . Obsidian Tachylite (basic glass)	
Isotropic	Crystals and crystalline grains	<i>Fluorite</i> . (111), 109° <i>Leucite</i> (anomalous) <i>Analcime</i> anomalous (100) <i>Cristobalite</i> anomalous, scaly <i>Sodalite</i> . <i>Hauyne</i> Garnet	
Uniaxial		Bright interference colours	Low interference colours
	Crystals, grains, granular aggregates	Scapolite, t. — (110) Zircon*, t. + Calcite ¹ , h. — (1011), d. r. 0.18 Cancrinite, h. —, d. r. 0.03	Melilite, t. — (110) <i>Leucite</i> , anomalous Quartz, h. + Nepheline, h. — Apatite, h. —
	Pseudohexagonal lamellae	Talc (001), m. —	
Biaxial	Crystals, grains, granular aggregates	Straight extinction	Zoisite, o. + Andalusite, o. — Cordierite, o. — Enstatite, o. + (110) 92' (100)
		Oblique extinction	Orthoclase, m. (001) and (010) — Plagioclases, tr. \pm (001) — (010) = 92° Polysynthetic twinning Clinzoisite, m. + <i>Cristobalite</i> , —, scaly
		Kyanite, tr. — Wollastonite, m. — Tremolite, m. (110) 124° Diopside, m. + (110) 87° (Plagioclases)	

Table 8

the Most Important Rock-Forming Minerals

Blue and violet	Green, yellow, brown, red (pink)	
—	<i>Palagonite. Tachylite. Limonite*</i>	
Fluorite Hauyne Sodalite *	Garnet. Rhombic dodecahedra Spinel. Octahedra Chromite *. Octahedra, semi-transparent	
	Non-pleochroic	Pleochroic
Anatase. t. — Corundum, h. — (1011) Tourmaline, h. —	Rutile*, t. + Haematite*, h., trans- lucent, red (Zircon*) (Nepheline) (Apatite)	Rutile*, t. + Anatase*, t. — Tourmaline, h. —
		Biotite, m. —, strong d. r. Penninite, m. —, weak d. r.
Cordierite, o. —	Fayalite, o. — Andalusite, o. — (110)	Staurolite, o. + Hypersthene, o. — (110) 88° Anthophyllite, o. — (110) 124°
Kyanite, tr. — Glaucophane, m. — (110) 124° Titanaugite, m. + (110) 87°	Augite, m. + (110) 87° Diallage, m. + (110) 87° (100) Sphene, m. + (Leuco- xene)	Aegirine, m. — (110) 87° Hornblende, m. — (110) 124° Epidote, m. — Titanite, m. +

	Colourless (sometimes grey, turbid)	
Tablets, plates, tabular aggregates	Talc, m. — Muscovite, m. — Prehnite, o. +	<i>Tridymite</i> , + Antigorite, o. —
Radiating and fibrous aggregates	<i>Thomsonite</i> , o. + Aragonite, o. — (Natrolite)	<i>Natrolite</i> , o. + Chalcedony, o. — Chrysotile, o. + <i>Scolecite</i> , m. —

1. Calcite often gives dull interference tints of the highest orders, which are gypsum plate.

it is converted into the higher volume γ -form, causing the slag to shatter into powder.

The characteristic features of some of the most important rock-forming minerals are given in Table 8, which can be used to make an approximate determination of them from the results of a microscopic investigation.

When a thin section is examined under the microscope, attention should first be paid to the colour of the mineral in simple transmitted light. This feature determines in which vertical column of the table the mineral in question must be sought. For coloured anisotropic minerals, each column is divided into two parts: in one, the minerals are pleochroic, and in the other do not possess pleochroism; minerals in which the pleochroism is very feeble, or sometimes either unnoticeable or doubtful, are placed in both columns. For colourless anisotropic minerals, there are two vertical columns: bright polarizing colours and dull, bluish-grey colours (as in apatite).

Another easily perceptible feature is the external outline of the grain being studied and this determines the horizontal grouping. The mineral sought will be one of those printed in a compartment at the intersection of the relevant horizontal and vertical divisions.

The system to which a mineral belongs is indicated by a letter to the right of each (c.—cubic, t.—tetragonal, h.—hexagonal, o.—orthorhombic, m.—monoclinic, tr.—triclinic). The sign of the optical character follows the system: plus (+) for positive, and the minus sign (—) for negative.

Minerals with negative relief are distinguished from the others by italic print (*leucite*, *tridymite*, etc.); minerals with indistinct

Table 8 (continued)

Blue and violet	Green, yellow, brown, red (pink)	
		Biotite, m. --, strong d.r. Chlorite, m. —, weak d.r.
		Chrysotile, o. +

distinguished from colours of the first order by their insensitivity to the first-order

or weak positive relief (nepheline, quartz, plagioclases) in ordinary print; thick letters characterize minerals of sharp relief (olivine, pyroxene); when the relief is extremely sharp (titanite), the thick print is accompanied by an asterisk after the name of the mineral.

If the mineral is characterized by some very typical cleavage, or if it is distinguished by this feature from the neighbouring minerals in a series, then this is shown, the symbol indicating the plane along which the cleavage takes place, and the angle in the case of prismatic cleavage.

CHAPTER VIII

Igneous (magmatic) rocks

The forms in which magmatic rocks occur are known from the course in geology, therefore only some of the more widely employed terms will be given here.

The terms *deep-seated* (or intrusive) and *effusive* are used depending on whether the solidification of the magma took place below the earth's surface or on it.

Intrusive rocks can form bodies of the very different dimensions in the crust. Large bodies (tens of kilometres in area) are called *batholiths*. Batholiths were originally supposed to be protrusions of magma into the crust and to be bottomless. It has recently been shown, however, that underlying rocks can be found for them also; relatively smaller intrusive bodies bear the name massifs, and those still smaller, *stocks*. There are frequent occurrences of small intrusions irrupting into sedimentary rocks along the bedding (for example, *laccoliths*). The magma here raises the overlying beds of sediments a little, and is injected between them, forming a loaf-shaped body; naturally, this "loaf" must have a feeder channel. Laccoliths may be simple, if there is only one intrusive body, or complex, if feeder channels leave it and, in their turn, form new and smaller laccoliths. If the irruption of the magma causes a depression in the underlying sediments, the intrusion becomes bowl shaped, and is then called a *lopolith*.

Flattened bodies of various sizes occur along with the more or less equidimensional magmatic bodies, and are usually termed *veins* or *dykes*. Veins can be *discordant* or *concordant*. The latter are injected along the bedding planes of the enclosing sediments; concordant veins are sometimes termed *sills* in distinction to the discordant ones, for which the accepted name is dykes. Branches from veins or from any other intrusive body are called *apophyses*.

Effusive rocks occur in the form of flows, sheets or domes. The characteristic feature of a flow is the considerable excess of length over width and thickness; sheets are sometimes of immense size,

but of relatively small thickness, and very great width and length: domes are loaf-shaped bodies of great height and small horizontal dimensions, formed by viscous lavas only. Lava can also congeal in a feeder channel. If this channel has a considerable extent and small thickness, it can be called a vein. Equidimensional forms are called volcanic necks. A neck can be filled with coherent lava or with a lava breccia.

The textures of magmatic rocks. The cooling conditions of a magma are expressed in the size and shape of the crystals formed in the rock. During the rapid crystallization which takes place in the formation of effusive rocks, part of the lava has no time to crystallize and congeals in the form of a glass; besides this, many centres of crystallization make an appearance as embryonic crystals (what are called microlites), which, with the glass, are extremely characteristic of effusive rocks. It is otherwise when a magma crystallizes in the depths of the earth. Here the solidification proceeds slowly. The highly volatile substances (water, compounds of boron and fluorine) which give mobility are gradually removed from the magma, thereby assisting the crystallization; a relatively small amount of large crystals is formed in the rock, absolutely no glass remains, and the final structure is said to be holocrystalline.

The glassy textures of effusive rocks can be further subdivided into individual textures* according to the quantity of microlites and how they are arranged.

If a rock contains only glass and there are no microlites or very few, the texture is holovitrophyric; if microlites are present but in very small amounts, so that glass obviously predominates, the texture is vitrophyric.

If the contents of microlites and glass are approximately equal, or the microlites are a little more plentiful, giving the appearance of a felted mass set in glass, the texture is termed hyalopilitic (or andesitic).

If there is very little glass, and it is contained in the spaces between the microlites, the texture is termed intersertal, and is closely related to the doleritic (diabase) texture, in which the interstices between elongated crystals of plagioclase are filled not with glass, but with some such mineral as pyroxene or chlorite.

When all the microlites are drawn out in a certain direction, as distinct flows, the texture is called fluidal (the result of crystallizing while the lava flow is moving).

The effusive textures listed are those formed when crystallization takes place as one process, but this is not always so. Crystallization can begin while the lava is still in the neck of the volcano or in the

* The names given to the textures in what follows are mostly those accepted internationally. For reasons of the space available in this textbook, only the commonest textures are given.



Fig. 68. Porphyritic texture. Biotite porphyrite with phenocrysts of zoned plagioclase and decomposed biotite, $\times 20$

magma basin which supplies the lava to the point of eruption, and it will then proceed more slowly and under conditions similar to those under which intrusive rocks crystallize. The resulting crystals will naturally be large and well formed, very different from the microlites crystallizing when a lava congeals at the surface. Such crystals are called porphyritic phenocrysts, and the whole rock possesses porphyritic texture (Fig. 68); that part of the rock enclosing the phenocrysts is called the groundmass and can have any of the textures listed above. Intrusive rocks also can have porphyritic structure, but, unlike effusive rocks, they have a holocrystalline groundmass.

Textures in holocrystalline rocks can be distinguished according to the extent to which faces can be seen on the minerals forming the rock. If the crystal of any mineral is well formed and possesses its characteristic faces, it is said to be *idiomorphic* (possessing its own form); but if it is filling up the spaces between other minerals, and its boundaries reflect the form of the other minerals, it is *allotriomorphic* (possessing the form of something else).

Of the great number of textures appertaining to the intrusive rocks, the following must be noted as being the most common.

If the rock is entirely composed of idiomorphic crystals, we speak of a *panidiomorphic* structure.

The texture of a holocrystalline rock, where the idiomorphism of the constituent minerals is clearly seen to be relative but not complete, bears the name *hypidiomorphic-granular*.

It is sometimes called granitic, but this is not correct, since a similar structure is also found in rocks other than granites.

If all the grains are irregularly formed and do not have their own boundaries, the texture is called *panallotriomorphic*. If the grains of some minerals are very large and enclose small grains of other mine-

erals without any kind of orientation, then the texture is termed *poikilitic*. When fine grains of one mineral are enclosed in large grains of another, and all the small grains have the same orientation, and appear as if they were parts of a second large crystal intergrown with the first, the structure is called pegmatitic (or granophyric).

Such an intergrowth is most frequently encountered in the case of orthoclase and quartz.

The mineral and chemical composition is closely connected with the composition of the magma from which the rocks have been formed.

It is important to note that known maximum and minimum contents of any one constituent exist for each group of rocks, and some combinations of these constituents are usual, others being impossible; a complete parallelism is seen between the chemical composition of deep-seated and effusive rocks; all chemical types of igneous rocks can be found in the deposits of any geological period.

Various types of deep-seated rocks can sometimes be observed within the confines of one intrusive massif, and one volcano can produce lavas of different chemical composition during its period of activity. The number of parental magmas can be supposed to be very much less than the number of igneous rocks, in other words, one magma in the process of congealing can give rise to several types of rock differing in composition. The cause of this appears to be either the complex and diverse physico-chemical processes involved in the splitting up of the magma itself, i.e., in its differentiation (in the first place, the separation due to specific gravity or gravitational differentiation), or the absorption of the rocks surrounding it—assimilation; sometimes, both processes can proceed in parallel.

It is thought that the composition of the original magmas could be arrived at by obtaining still more data concerning the physical chemistry of the processes taking place in the magma, and by studying the distribution in the crust of the different magmatic rocks which are the products of these processes.

Statistical calculations have shown that the distribution in nature of rocks of different composition is very unequal. This is shown as a graph in Fig. 69. All the rocks are grouped according to their SiO_2 content, this content being plotted along the horizontal axis. The area occupied by a rock of given composition is plotted along the vertical axis. The graph shows that the extreme types containing little SiO_2 (less than 35%) and those quite rich in it are found extraordinarily rarely; intermediate types occur much more often, those most commonly found being rocks containing 72.5% of silica (as in typical granites) and 52.5% SiO_2 (as in typical basalts and gabbros). The distribution of the intermediate rocks is less. Both maxima are sharper still, if the intrusive and extrusive rocks are considered separately.

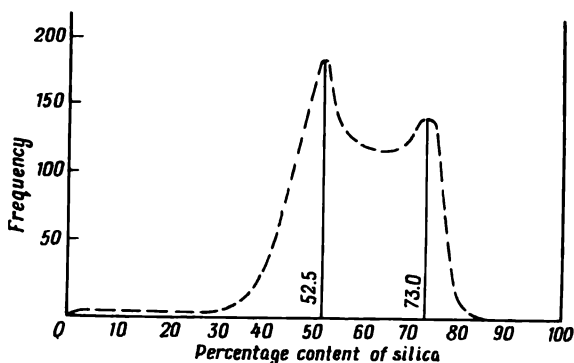


Fig. 69. Distribution of igneous rocks of different composition

Of the intrusive rocks, granites and granodiorites cover an area of the earth's surface which is 20 times greater than that covered by all the remaining rocks taken together. Of the effusive rocks, the basalts and the pyroxene andesites closely related to them in composition are at least 50 times greater in volume than all the remaining rocks taken together.

It has been suggested that there were originally only two parental magmas in the crust, an acid (granite) and a basic (basaltic), and that rocks of extreme composition were formed as a result of their differentiation, while a mixing of these magmas gave the intermediate varieties.

For all its attraction, such a scheme encounters objections, which are subsequently discussed in more detail. Many specialists, for example, consider that there was only one parental basaltic magma, and that a granitic magma was obtained from it by differentiation. The authors of this view find their main evidence in the fact, that during volcanic eruptions, when the magma proceeds directly from a magmatic focus and has no time to undergo differentiation, it pours out and congeals as basalt. Consequently, it is this rock which must be nearest in composition to the parental magma. But if the congealing takes place in depth, all the conditions necessary for the differentiation of the magma and the formation of the low melting granitic rocks are available.

Independently of whether we shall consider that there are one or two original magmas, there does exist in nature a sharp division of all magmatic rocks into two large groups according to composition, and customarily designated acid and basic.

The *acid* rocks are characterized by excess silica, by the absence of minerals unsaturated in silica, and by containing only very little iron and magnesium.

Minerals containing fluorine and boron are often associated with these rocks, as also are numerous deposits of non-ferrous metals,

gold, tinstone, bismuth, molybdenum, tungsten and other ores.

The *basic* rocks are characterized by a deficiency of silica and, associated with this, the presence of a more or less significant quantity of silicates unsaturated in SiO_2 . They contain a lot of iron and magnesium. Sometimes iron, in the form of magnetic iron ore (or titanmagnetite), is found in such quantities that it can be regarded as an ore. Deposits of iron, chromium, copper, nickel and cobalt, as well as platinum and metals of the platinum group, are associated with these rocks.

The igneous rocks can be divided into five groups according to their relative content of SiO_2 :

ultra-acid rocks—strongly supersaturated in silica (SiO_2 greater than 75%);

acid—supersaturated in silica (SiO_2 from 75 to 65%);

intermediate—saturated in silica (SiO_2 from 65 to 52%);

basic—unsaturated in silica (SiO_2 from 52 to 40%);

ultrabasic—strongly unsaturated in silica (SiO_2 less than 40%).

Apart from silica, the following enter the composition of magmatic rocks as essential constituents: Al_2O_3 , Fe_2O_3 , FeO , MgO , CaO , K_2O , Na_2O , H_2O and certain other substances, but only the alkaline earths and the alkalis are of any importance for classification. The first are characteristic of basic rocks, and the second of acid. There is also a special group of alkaline rocks.

The mineral composition of rocks is bound up with their chemical composition. The following are the most important mineralogical features for the groups of rocks just given:

Ultra-acid rocks	Quartz is exceptionally abundant
Acid rocks	Quartz is the main constituent
Intermediate rocks	Neither quartz nor basic plagioclase
Basic rocks	Characterized by basic plagioclases, olivine may be present
Ultrabasic rocks	No plagioclases, coloured minerals only present: olivine, pyroxene, etc.

The presence of feldspathoids, with or without feldspars, is characteristic of *alkaline* rocks. Because of their SiO_2 content, these rocks can belong only to intermediate or basic rocks.

CHAPTER IX

Causes of the diversity [of magmatic rocks]

It has already been pointed out that when an intrusive body solidifies, a series of physico-chemical processes take place and lead to the formation of various rock types. The details of these processes are not yet completely understood.

The hypotheses discussed in this chapter are those which have received the most acceptance in recent years.

1. MAGMATIC DIFFERENTIATION

According to Lewinson-Lessing, a magma can split into two immiscible liquids (liquation of magma) as the result of cooling or loss of volatiles, one of these liquids forming an emulsion in the other.

The author of this hypothesis sees evidence for the existence of such magmatic emulsions in the existence of variolites and banded gabbro.

The variolites are effusive rocks of a special structure; their fine-grained or glassy groundmass encloses rounded areas (varioles) of another chemical and mineral composition. The varioles are claimed to be the relicts of the emulsion, or disperse phase of a dissociated primary magma.

Banded gabbros are characterized by interbedded layers enriched in dark minerals, with bands (layers) of concentrations of light minerals, plagioclases, for example. Lewinson-Lessing suggested that these bands are the result of liquation of the magma, both constituents being attenuated in the course of congealation.

It is also quite probable that the magmas created by liquation could separate from each other under the effect of different specific gravity, and freeze separately as different rocks.

It is also possible that chemical differences can appear in various parts of the magma basin without liquation, and merely as the result of differences of temperature and pressure.

2. CRYSTALLIZATION DIFFERENTIATION

Most crystals formed in the first phase of crystallization of a magma are heavier than the enclosing magma, and therefore tend to sink. The amount of crystals sinking depends on their density and shape, and the viscosity of the residual liquid. Numerous cases are known of accumulations of magnetite, pyrrhotite, olivine and other heavy minerals near the base of an intrusive body, their presence being explained by crystallization differentiation.

If this deposit of crystals is again subjected to solution, new magmas will be formed and will differ in composition from the primary magma.

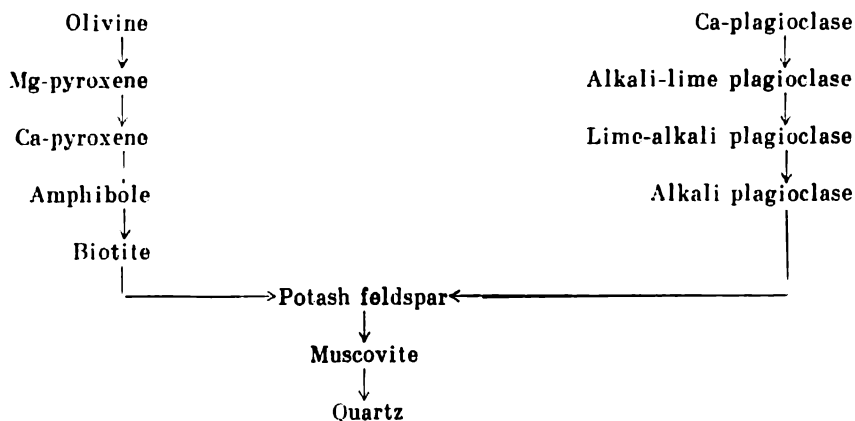
When minerals which have separated from the melt remain in it without any migration due to specific gravity, and if they form an isomorphous mixture, they will continue to react with the melt, and their composition will change continuously as the intrusion cools.

There is another type of reaction possible between melt and crystals: at a certain temperature, an earlier formed mineral may begin to dissolve in the melt, and a new mineral is formed at the expense of the first; thus olivine separates first in some andesites, then reacts with the melt to form pyroxene.

The crystallization of a magma is thus known to be an evolutionary process and the composition of the magma in the initial phases is absolutely different from that obtained as the result of the complete solidification of the rock.

These relationships can be represented schematically (Bowen's reaction series).

In this scheme, each successive mineral participates in the creation of the next mineral down, either by reacting with the residual melt, or by direct conversion. If the first phases are isolated, an almost



monomineralic magma will, according to Bowen, be obtained (dunites and suchlike), while in the later phase of the crystallization, gabbros and syenites will be formed, and finally, as the end product of the differentiation, a granite magma with its pegmatites and quartz veins.

3. ASSIMILATION

The phenomenon of *assimilation* plays an important part in the formation of the different igneous rocks; it involves the engulfment and melting, or contact recrystallization of rocks with which the magma is in contact during its injection into the crustal layers. The result is an alteration in the composition of the magma which leads to the formation of new rock types, contact or hybrid rocks, sometimes very different from the magma which gave rise to them.

The presence of cordierite, sillimanite, garnet and staurolite in magmatic rocks, for example, granites, is a sign of the assimilation of argillaceous sedimentary rocks. Diopside and calcium silicates appear in magmatic rocks when limestones have been assimilated.

It can happen that a magma engulfs fragments of its enclosing rock, and that these fragments are not wholly digested. Fragments which have been partially melted or strongly altered at their contact with the magma, and sometimes even fresh extraneous fragments in the magma, are called *xenoliths*. They are of especial interest to research workers because they allow the conditions under which the magma was irrupted to be more precisely known.

* * *

The theory of magmatic differentiation made its appearance in petrography earlier than others on the basis of laboratory experiments at the beginning of the twentieth century, and has been increasingly opposed (but cannot be considered to have been completely discarded) since natural magmatic melts contain volatile components which are always absent in laboratory melts and whose presence can substantially alter the behaviour of a silicate liquid.

The theory of crystallization differentiation was put forward later, in the second decade of the twentieth century, and based wholly on laboratory data. It was at first very attractive, but in the course of time, many objections piled up against it, partly from laboratory experiments, but particularly from physico-chemical interpretation of natural phenomena.

Gravitation differentiation, assimilation and hybridism are now given pride of place by the majority of petrographers when accounting for the diversity of magmatic rocks. These views are discussed below in greater detail, after acquaintance has been made with the metamorphic rocks.

CHAPTER X

Classification and nomenclature of rocks

Present-day classifications of rocks are based on genetical features, rocks being first of all distinguished as either intrusive or effusive. Sometimes a special group of vein rocks is also separated, but these are more often considered with those rocks to which they are nearest in structural peculiarities, genesis and composition.

In what follows, the division of rocks will be according to chemical or mineralogical features.

Originally, attention was mainly paid to chemical classifications, but recently more and more consideration has been given to quantitative mineralogical considerations.

When comparing the chemistry of rocks and rock-forming minerals, a number of complications are often encountered, caused by isomorphism, solid solutions, and the presence of the finest of mechanical mixtures in certain minerals. Besides this, some chemical compounds are known to be capable of forming different minerals under different physico-chemical conditions. In such cases, it is useful to compare the total chemical composition of the rock with a quantitative determination of its mineralogical composition under the microscope. But even this is frequently incapable of resolving all uncertainty. Separate chemical analyses should then be carried out on minerals which manifest any sort of optical anomaly. Spectroscopy is being more and more used for this, being rapid and accurate enough for the purpose.

The composition of every mineral in a rock can now be determined adequately by microscopical methods of investigation, and also its approximate amount; besides this, the mineralogical study of a rock can be carried out so much more quickly under the microscope than a detailed chemical investigation, with the result that there is a tendency to give preference to mineralogical classifications. However, they do not exclude chemical methods of investigation, which have not lost their importance in petrography, but, on the contrary, make it possible to come nearer an objective solution of the most complex

A Mineralogical Classification

Feldspathic rocks with pyrox

Texture	Alkali feldspar only (orthoclase, anorthoclase, sanidine, microcline, albite)		Alkali feldspars predominate, but soda-lime feldspars also present		Alkali and soda-lime feldspars (orthoclase and plagioclase) approximately equal	
	with quartz	without quartz	with quartz	without quartz	with quartz	without quartz
Granitoid	Alkali granites	Alkali syenites	Granites	Syenites	Adamellites	Gabbrosyenites gabbro-norites Syenite-diorites
Granite-porphyrific	Alkali granite porphyries	Alkali syenite porphyries	Granite porphyries	Syenite porphyries	Grano-diorite porphyries	Monzonite porphyries
Porphyritic or glassy	Older Types	Alkali quartz porphyries Quartz keratophyres	Keratophyres	Quartz porphyries	Orthoclase porphyries	Quartz porphyries
	Newer Types	Alkali liparites	Alkali trachytes	Liparites Trachyliparites	Trachytes	Dacites Liparite-dacites

Table 9

of Igneous Rocks

nes, amphiboles or mica

Feldspars — soda-lime plagioclases (sometimes sporadic orthoclase)				Without coloured constituents	
with quartz	without quartz			with quartz	without quartz
	with acid plagioclase	intermediate	with basic plagioclase		
Quartz diorites	Diorite	Gabbro-diorites	Gabbro Gabbro-diabases	Alaskite	Anorthosite Labradorite
Quartz-diorite porphyrites	Diorite porphyrites		Gabbro-porphyrite Diabase porphyrites	Aplite	
Quartz porphyrites	Porphyrites		Diabases Augite porphyrites Spilites Melaphyres		
(Dacites) Andesite-dacites	Andesites	Andesite-basalts	Basalts		

Table 10

A Mineralogical Classification of Igneous Rocks

Texture	Feldspar-feldspathoid rocks			Feldspathoid rocks	Rocks without feldspar		
	with orthoclase (sanidine)	with orthoclase and plagioclase	with plagioclase		with quartz	without quartz	without quartz
Granitoid	Nepheline syenites Leucite syenites		Teschenites	Ijolite Urtite	Greisen	Pyroxenites	Peridotites Dunite
Granito-porphyrific	Nepheline and leucite porphyries						Picrites
Porphyrific or glassy	Older Types	Phonolites	Melaphyres				
	Newer Types	Phonolites	Trachy-basalts	Leucitophyre			

questions of the genesis of rocks, especially in connection with new investigations in the field of physical chemistry.

Some of the most important classifications now employed in petrography will be discussed in this chapter.

1. QUALITATIVE MINERALOGICAL CLASSIFICATION

The simplest mineralogical classification of rocks is presented in Tables 9 and 10.

Rocks of the same composition are divided according to texture into intrusive (equigranular and porphyritic) and effusive; the latter are divided into Older Types, having the appearance of ancient rocks, and Newer Types, having the appearance of young rocks. The Older Types are usually decomposed, the glass in them often replaced by a quartz-feldspar (felsitic) mass in acid rocks, and by chloritic material in basic varieties.

A more detailed determination of a rock is based on the quantitative relationships of the minerals and their character, the number of the plagioclase, the composition of the olivine and the character of the pyroxene being of great importance.

2. QUANTITATIVE MINERALOGICAL CLASSIFICATION

In many cases, it is impossible to name a rock accurately from the qualitative mineral composition, and so classifications based on a calculation of the percentage content of its constituent minerals have come into use.

The most rational attempt at a classification of this kind is that of the granitoids, proposed by B. Kupletskii.

Since four components—quartz, orthoclase, plagioclase and coloured and accessory minerals—enter into the composition of the granitoids, the system can be represented by a three-dimensional diagram in the form of a tetrahedron, with these groups of minerals at its apices.

Kupletskii suggested that this could be represented on a plane by taking five sections of this tetrahedron (Fig. 70), the ratio of *K*-feldspar to the total feldspar in the rock being taken as the basic coordinate for the choice of section; then quartz (*qu*), feldspar (*fel*), coloured and accessory minerals (*col*) will be situated at the apices of the triangular section of the tetrahedron. In section *I*, the rocks represented contain 0-10% of *K*-feldspar in their total feldspar content; in section *II*, 10-30%; in section *III*, 30-60%; in section *IV*, 60-80%; in section *V*, 80-100%.

In each section of the tetrahedron, the rocks are further subdivided according to their quartz content. The percentage composition of

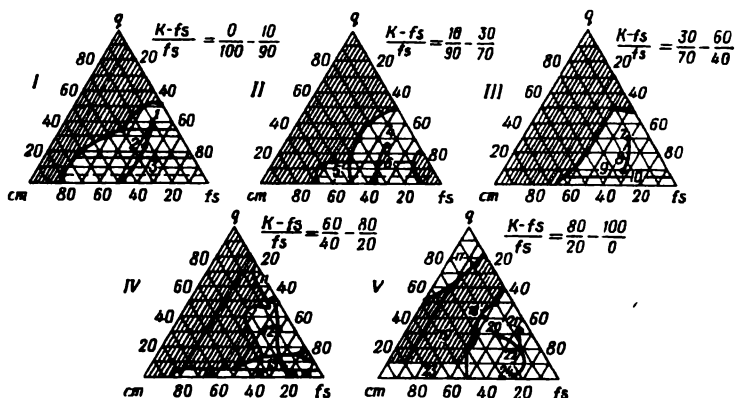


Fig. 70. Kupletskii's classification of the granitoids:

1—plagiogranites; 2—granodiorites; 3—diorites and quartz diorites; 4—granodiorites; 5—diorites; 6—diorites and monzonites; 7—granites; 8, 9—adamellitites and quartz monzonites; 10—monzonites and syenite-diorites; 11—granites; 12—granites; 13—quartz syenites; 14—shonkinites; 15—monzonites; 16—syenites; 17—greisen; 18, 19, 20—alkali granites; 21—alaskites; 22—quartz syenites and granosyenites; 23—shonkinites; 24—syenites

K-feldspar in the total feldspar as estimated in thin section determines the section to which the rock under examination is to be referred. For example, let the rock contain 25% of quartz, 40% of oligoclase, 30% of K-feldspar and 5% of coloured minerals. The ratio of K-feldspar to the total amount of feldspars is 43% ($30 : 70 = 0.43$), i.e., the rock is situated in section *III*, and according to its coefficients (25% of quartz, 70% of feldspar, 5% of coloured minerals) falls in the field of granites and is named accordingly.

Similar diagrams have been worked out by Kupletskii for other deep-seated (intrusive) rocks, thus allowing them to be named with great precision if their quantitative mineralogical composition is known.

The situation with regard to the effusive rocks is not so good. Here a considerable amount of glass is usually present and indefinite in composition, so that the rock cannot be determined by a calculation of the quantitative mineralogical composition. The composition of the glass must therefore be determined as well. In the simplest cases, this can be arrived at by using the immersion method to measure the refractive index of the glass. In acid rocks, the glass is rich in silica and possesses low refraction ($n = 1.49-1.51$). In basic rocks, the glass is rich in iron and magnesium and has a high refraction ($n = 1.56-1.60$). But the water content and other factors affect the refraction of a glass, and so some uncertainty is introduced into any estimate of its composition. Nevertheless, the method of rock determination by calculation of the mineral composition is the most convenient and simple.

CHAPTER XI

Description of magmatic rocks

Only the most important of the magmatic rocks will be discussed in this chapter, the intrusive representatives of any one petrographic group being dealt with first, and then their effusive analogues.

The description begins with the acid rocks, the most widespread on the earth's surface; a review is then given of the alkaline, intermediate, basic and, finally, the ultrabasic rocks, but it must be borne in mind that transitional varieties can always be found.

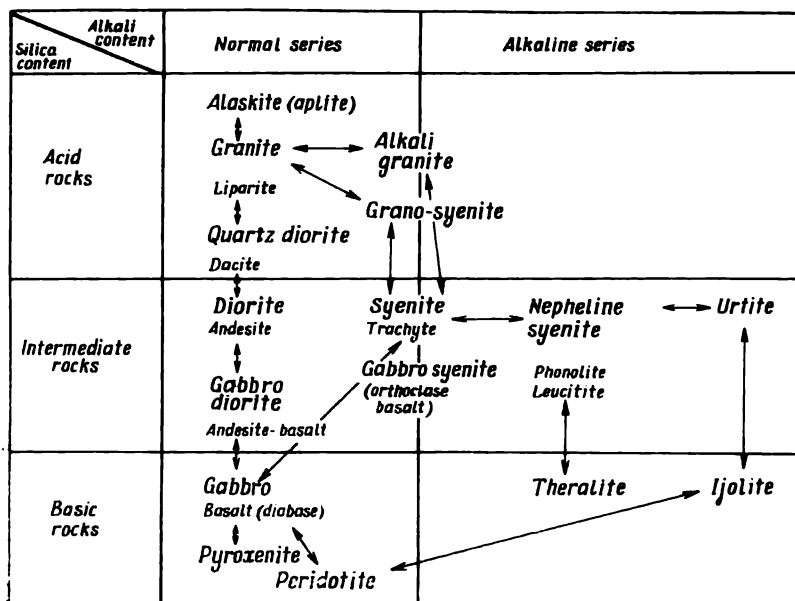


Fig. 71. The mutual relationships of the most important rocks

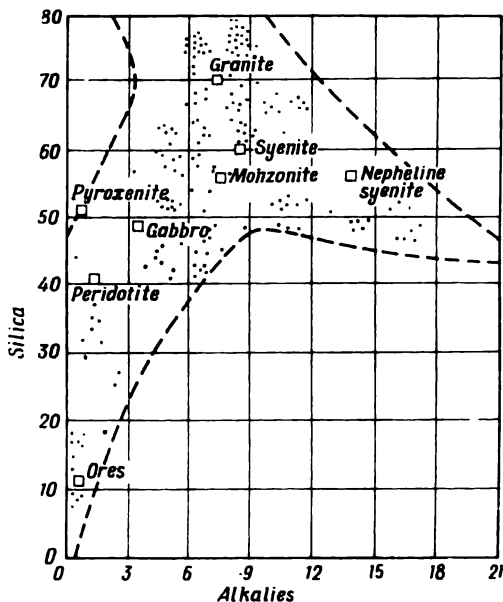


Fig. 72. Diagram of the compositions of the most important rocks

The mutual relationships can be represented schematically as given in Fig. 71 (Roman type—intrusive, italic—effusive rocks; vein rocks are given in brackets).

The diagram shown in Fig. 72 has been obtained by using rectangular coordinates and plotting the amount of silica along one axis and the amount of alkalies along the other; the rock types are indicated by little squares, and the individual analyses by points, the whole field of magmatic rock analyses being outlined by broken lines.

1. INTRUSIVE ACID ROCKS

The group of the granites belongs to the intrusive acid rocks.

The granite group consists of crystalline granular rocks (see Fig. 20), the essential constituents of which are quartz, alkali feldspar, acid plagioclases and some dark mineral from the mica group or from the amphiboles and pyroxenes. Biotite or muscovite can be a constituent, more rarely lithium mica; light and dark micas are often found together along with a pyroxene-augite, hypersthene or aegirine, and, more rarely, amphibole as either hornblende or riebeckite.

The alkali feldspar contained in granites is most often microcline. Orthoclase granites or granites with anorthoclase are also common; the plagioclase is of the oligoclase series, and, more rarely, albite.

On decomposition, feldspars give finely dispersed kaolinite or very small flakes of mica (sericite).

Quartz usually forms irregular grains, except when it lines the walls of drusy cavities.

Biotite is typified by the presence of inclusions of apatite, rutile and magnetite, fine grains of zircon and other radioactive minerals which may be surrounded by a halo in which the mica is of considerably intenser colour and pleochroism (pleochroic haloes are the result of radioactive emanations). Biotite becomes green on decomposition, being converted to chlorite. Sometimes epidote and magnetite are formed along with chlorite.

It is only rarely that muscovite can form independent irregular flakes; it is most frequently found as an accessory or secondary mineral (sericite).

Hornblende forms irregular grains, sometimes twins on (100). Under the microscope, the colour is green, or a brownish-green with noticeable pleochroism. Usual decomposition products are a green chloritic material or epidote and quartz.

Pyroxenes, mainly augite or diopside, are colourless in section, and on decomposition, pass into hornblende, or break down into a green chloritic aggregate. Augite is sometimes accompanied by orthorhombic pyroxene (enstatite or hypersthene).

The accessory minerals found in granites include magnetite, haematite, pyrite; sphene (titanite) as angular or rounded grains of grey-brown colour; rutile as tetragonal prisms, sometimes elongated, or as geniculate twins, often as an inclusion in quartz and feldspar; zircon with characteristically bright interference tints and short tetragonal prisms; apatite as narrow, hexagonal, colourless prisms. Minute reddish garnets are often found in some muscovite granites and aprites.

Tourmaline is frequently found in granites, especially near the edge of a massif, occurring as well formed crystals (trigonal prisms), or, more often, as irregular grains of brown colour, sometimes with pale blue patches.

The normal texture of granites is crystalline granular (hypidiomorphic-granular), i.e., a texture which shows a gradual decrease of idiomorphism, the minority of the crystals being idiomorphic from the start to finish of the crystallization. An equigranular texture can occur for different grain sizes (coarse-, medium- or fine-grained) or porphyroidal (granite porphyries). When the mica or secondary chlorite in granite, subjected to metamorphism, is collected into bands, the structure is similar to that of gneiss (granite gneisses, gneissose granites). Sometimes pegmatite structure is found, occasioned by regularly oriented intergrowths of quartz and coarse crystals of potash feldspar.

As the plagioclase content in a granite increases, a transition is observed to a variety in which orthoclase and plagioclase play an

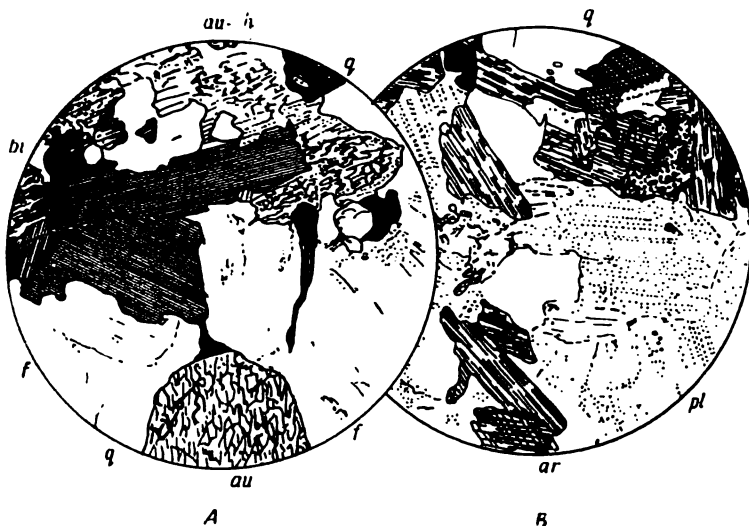


Fig. 73. Quartz diorites, $\times 20$:

A—the feldspar (f) is sharply zoned between crossed nicols but is near oligoclase in composition. The ferromagnesian minerals are biotite (bi) and colourless auzite (au) which is sometimes converted into a light yellow hornblende (au-h). Isolated grains of magnetite, apatite and sphene; quartz (q) was the last to crystallize. B—granodiorite, consisting of green hornblende and plagioclases (pl) with a certain amount of biotite, orthoclase (or), quartz (q), sphene, magnetite and apatite

equal part. These rocks are adamellites. With a still greater content of plagioclase, granodiorites are obtained, forming a transition to the quartz diorite family, in which feldspar is represented exclusively by acid plagioclase, sometimes with a negligible amount of orthoclase (Fig. 73). Still other intermediate varieties are possible. Lewinson-Lessing proposed that these and similar rocks which are difficult to distinguish from one another in the field should be placed together under the general name of *granitoids*.

Special notice should be made of the feldspar-free granite rock which is an aggregate of mica and quartz, and known as *greisen*.

Alaskite is a rock very near *greisen**, often forming large massifs and consisting of alkali feldspar and quartz without coloured minerals or with only an insignificant amount of them. Similar rocks are fine-grained granites or *aplites*, very poor in mica, or containing only muscovite, and usually occurring as veins.

* The term "greisen" (according to Lewinson-Lessing) signifies granitoid magmatic rocks richest in dark minerals, "ultra-melanocratic granitoids". However, this term is also used to designate metamorphic rock of quartz-mica composition; it can be a metamorphosed granitoid, as well as different silicate rocks of another nature. Metamorphic greisens should be clearly distinguished from magmatic greisen. Metamorphic greisens are usually ore-bearing.

Two great groups of granites are distinguished, depending on composition. Firstly, *normal granites*, rich in alkalies and lime with potash feldspar and acid plagioclases developing together. Biotite is usually the only coloured mineral present. Secondly, there is the group of the *alkali granites* and the *granitoids*, in which CaO is almost completely absent, and Na₂O predominates. Here almost pure albite is present alongside with potash feldspar in large quantity; there are no plagioclases. Soda pyroxenes and amphiboles play a major role.

Granites form intrusive bodies of various forms. These are often batholiths emerging at the earth's surface as huge massifs occupying areas of many square kilometres. Composition, texture and colour of the rock are maintained without change over great distances, areas of different composition only being observed locally. In some cases, the formation of these areas has been brought about by the presence of a large amount of xenoliths, in others, it is explained by peculiarities of the differentiation process.

Aplites and pegmatites are often found as veins in granites. Aplites are fine-grained rocks with panidiomorphic texture. Their mineral constituents are microcline, oligoclase, quartz, muscovite, red garnet or topaz, and colourless or pale tourmaline. Pegmatites are characterized by an extraordinary coarseness of grain, frequently by peculiar intergrowths of different minerals with one another, and peculiarities of composition. Pegmatites are usually formed of quartz, potash feldspars, acid plagioclase and muscovite; rare minerals are frequently present (minerals with rare earths, boron, fluorine, etc.) and as a result they are sources not only of mica and the raw materials of the ceramic industry, but of some rare elements and coloured stones (Figs. 74 and 75).

The origin of pegmatites is one of the important problems of modern petrography. Most investigators are of the opinion that during the congealing of the magma, a residue separated off much enriched in volatile substances. When injected into fissures in parts of the intrusion which had solidified earlier, or in the adjacent rocks, this pegmatite magma began to crystallize; the rocks first separated being very similar to aplites, and then frequent and very coarse crystals of feldspar and quartz began to form in the centre of the vein. It is here that crystals of beryl, topaz and other rare minerals containing volatile substances are encountered.

Zavaritskii refutes the presence of a special pegmatite magma or melt, and considers that pegmatites are formed as a result of the recrystallization of rocks under the action of a gaseous solution, which carries ore elements out of the congealing magma, and moves away from the massif of congealing rock towards the surface to form the various types of ore veins and lodes.

Granitoid rocks are usually intersected by numerous fissures, arising as a result of contraction in volume when the intrusion was cooling, and of various tectonic movements. The total effect of the cracks



Fig. 74. Pegmatite veins in granite

creates jointing. Parallelepiped, or coarse-bedded, so-called pillow jointing is characteristic of granites (Figs. 76 and 77). Granites, especially those possessing coarse jointing, can be used as building material for walls, foundations of bridges and other structures, as well as for the external facings of buildings. Sometimes the joints are so widely spaced that it is possible to obtain large monoliths without any cracks (for example, the base of the monument to Peter the Great, and the Alexander Column in Leningrad).

The greatest number of outcrops of granitic rocks in the U.S.S.R. are found in exposures on the ancient surface of pre-Cambrian formations in the Kola Peninsula and in the Ukraine.

In the central part of the Soviet Union, ancient granites outcrop in the Voronezh region; they are also found in the Urals, in the Baikal, Sayany and Ussurii regions. Still younger intrusions of granitic rocks are known. Numerous relatively small massifs are found injected into the carboniferous in the Eastern Urals, and a series of ore deposits in the Urals are connected with them.



Fig. 75. Inspecting a pegmatite vein (searching for crystals)



Fig. 76. Platy jointing in granite (the Central Caucasus)



Fig. 77. Pillow jointing in granite (the Katun Alps, Altai)

In the Caucasus, near Baikal and other regions of the Soviet Union, Jurassic and, in some cases, even Tertiary ore-bearing granites are known.

2. THE EFFUSIVE FACIES OF AN ACID MAGMA

The rocks included in this group are the *liparites*, newer type rocks (of Tertiary and post-Tertiary age) which are nearly or wholly fresh, and the *quartz porphyries*, which are altered Older Type (pre-Tertiary) rocks. Both rocks are near granite in composition. Quartz porphyries are usually of a red-brown colour, sometimes green; liparites are most often white, grey, light-yellow, sometimes reddish.

The texture of these rocks is porphyritic, the phenocrysts being crystals of quartz and sanidine in liparites, of quartz and orthoclase in quartz porphyries. The groundmass becomes dark to black in the presence of a large amount of glass. Sometimes, especially in quartz porphyries, the glass crystallizes as fine grains of quartz and feldspar, and the texture is then termed felsitic, and the whole rock is called a felsite, if no phenocrysts are present.

Glassy varieties with fresh glass are known under various names: *obsidian* is a volcanic glass with a vitreous lustre, and black or brown in colour (Fig. 78); *pitchstone* is a volcanic glass more or less rich in water and having a greasy lustre (Fig. 79); *perlite* is a glass with a series of concentric cracks along spherical surfaces (Fig. 80); *pumice* is a very vesicular glassy rock, which has been frothed up by a violent evolution of gases.

Effusive rocks, corresponding to the quartz diorites in chemical composition, bear the name of dacites if they are Newer Types, and quartz porphyrites, if they are Older Types. The phenocrysts in them are plagioclase, usually acid, about Nos. 20-25; only acid plagioclase is found as microlites. Quartz and orthoclase (sanidine) crystallize extremely rarely, occurring only when there is almost no glass in the rock; otherwise their constituent elements remain in the glass. Scaly cristobalite (spherulites) or spear-shaped crystals of tridymite are sometimes found in the vesicles of dacites.

3. ALKALINE ROCKS

The Syenite Group

Syenites are distinguished from granites by the absence of quartz (or by insignificant amounts of it) and by any proportion of the feldspathic and coloured constituents; they are richer in coloured constituents than granite. Amphibole, pyroxene and mica syenites are distinguished, the amphibole syenites being the most common.



Fig. 78. Obsidian

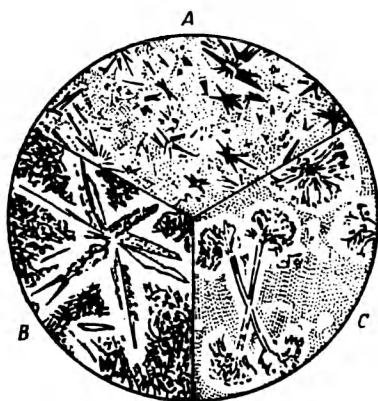


Fig. 79. Acid pitchstone:
A—asteriate grouping of dendritic crystallites, $\times 20$, *B*—the same, $\times 100$;
C—feathery crystallites grouped in crosses, $\times 100$

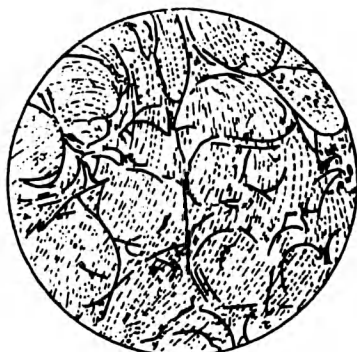


Fig. 80. Vitreous liparite (obsidian), $\times 20$. Sinuous flow lines are seen, intersected by a system of curved perlite cracks

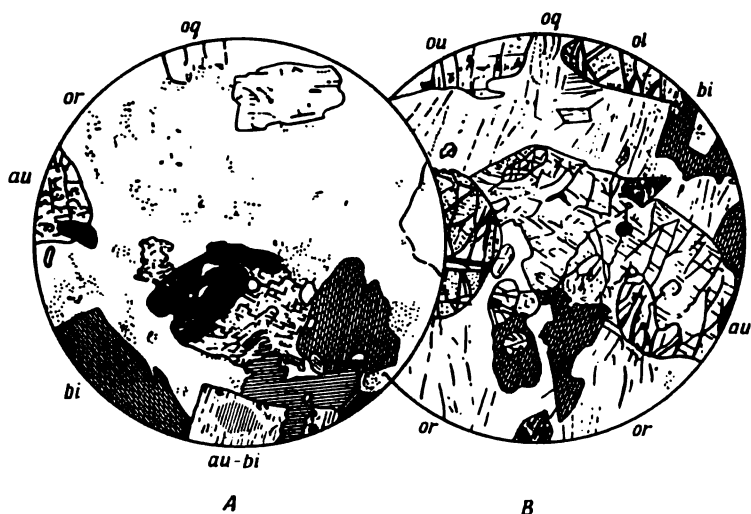


Fig. 81. Monzonites, $\times 20$:

A—typical monzonite, showing apatite, magnetite, colourless augite (au) and brown biotite (bi), sometimes intergrown oligoclase (og) and orthoclase (or); B—olivine monzonite. Ferromagnesian minerals in great amounts, with abundant olivine among them (ol).

Augite syenites containing basic plagioclase as well as orthoclase are called monzonites (gabbro syenites) (Fig. 81).

Syenites are also divided into alkaline syenites, which do not contain plagioclase, and normal syenites, with orthoclase and plagioclase. Syenite aplites and pegmatites occur.

The feldspars in syenites are often perthitic. The plagioclase is acid, most frequently oligoclase, more rarely oligoclase-albite or andesine, sometimes forming inclusions with other minerals in relatively large crystals of orthoclase. Biotite is usually present in platelets, sometimes earlier, sometimes later than the plagioclase.

The usual hornblende in syenites is mostly the same as in granites. Alkali syenites contain hornblende with intense brown absorption and pleochroism.

Pyroxene is represented in syenites by augite (Fig. 82) and diopside; the augite is sometimes rich in titanium, and it is a light brown or purple brown, with clear pleochroism; augite with a parting is found in some syenites. In alkali syenites, pyroxene is represented by bright green pleochroic aegirine or by pale green and less pleochroic aegirine-augite, a mineral intermediate between aegirine and common augite.

The biotite is dark chestnut-brown or green. In some alkali syenites it is extraordinarily rich in iron, and, as a consequence, the colour is so dark that vibrations parallel to the cleavage traces are almost completely absorbed, even in a slice of normal thickness.

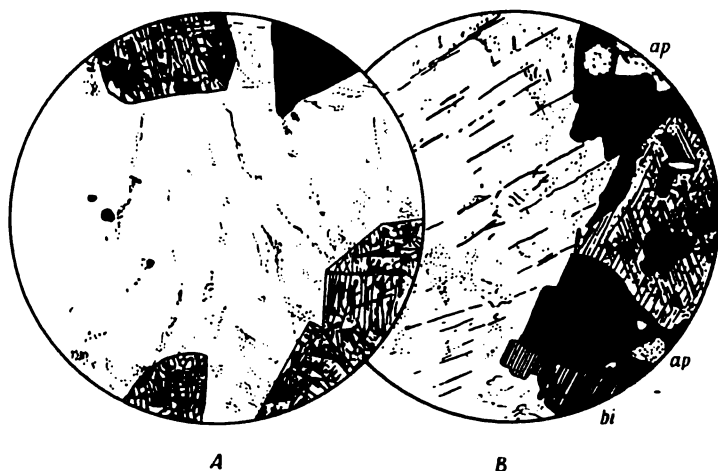


Fig. 82. Augite syenites, $\times 20$:

A—mainly alkali feldspars, idiomorphic light augite and opaque magnetite; *B*—brown augite, bordered with brown hornblende and biotite (*bi*), feldspar in large crystals, apatite (*ap*)

As accessory minerals, most syenites contain good crystals of sphene showing cleavage and often their characteristic twinning. Zircon, apatite, magnetite, ilmenite and haematite are also found in various amounts. Some alkali syenites contain dark chestnut-brown garnet.

Syenite bodies are usually markedly inferior in size to granites. They usually form stocks, localities and other forms of small massifs and are even found as veins.

The texture of syenites is hypidiomorphic-granular, but porphyritic varieties are found, also varieties with a gneissoid structure.

Syenites are very much less common than granites and play a relatively subordinate role. Syenites of the South Russian crystalline region, where they pass into gabbros, have been much studied, as have syenites of the Urals, which are associated with deposits of magnetic iron ore (the Hills of Blagodat, Vysokaya and Nizhny Tagil).

Trachytes and Orthophyres

Trachytes (Fig. 83) and their Paleovolcanic and Older Type analogues—the orthophyres (orthoclase porphyries) and keratophyres—are the effusive facies of the syenite magma.

Trachytes and orthoclase porphyries are characterized by the abundance of microlites, the absence of glass and the presence of flow lines around the phenocrysts—what is known as the trachytic texture. Trachytes and orthophyres often contain large phenocrysts of feldspar, but can also occur without them. Potash feldspar is the main mineral of the trachytes, and is usually sanidine, but in the ortho-



Fig. 83. Augite trachyte, $\times 20$. Sanidine phenocrysts, pale-green soda augite and magnetite. The groundmass shows fluidal texture

clase porphyries it is orthoclase, which occurs both as phenocrysts and microlites; dark minerals are extremely rare, biotite or hornblende, rarely augite, being found in trachytes, while in the orthophyres, the dark mineral is usually chloritized. Trachytes are pale in colour; they are often whitish grey, sometimes with a rough fracture as a consequence of a very fine porosity. Porphyries are darker, and are distinguished from trachytes in the same way as quartz porphyries are distinguished from liparites. They all occur in the form of flows or sheets, often accompanied by tuffs.

Special rocks, in which the feldspar is represented by albite, are grouped under the name of keratophyres; the dark minerals are chloritized as a rule, or altered in some other way; the glass has devitrified. Keratophyres are usually the product of submarine eruption.

The transitional varieties, trachy-liparites and trachy-dacites, occur.

Trachytes—rare in the U.S.S.R.—are found in the Caucasus and Transcaucasia; porphyries and keratophyres are found in the Urals, Siberia, the Central Caucasus and the Crimea. The trachy-liparite rocks of Beshtau are worked for acid-resistant material and building stone.

The Nepheline Syenite Group

Nepheline syenites. The distinguishing feature is the presence of nepheline and alkali feldspars. Plagioclases play a subordinate role in the majority of nepheline syenites, but the rocks known as nepheline monzonites contain almost as much basic plagioclase as ortho-

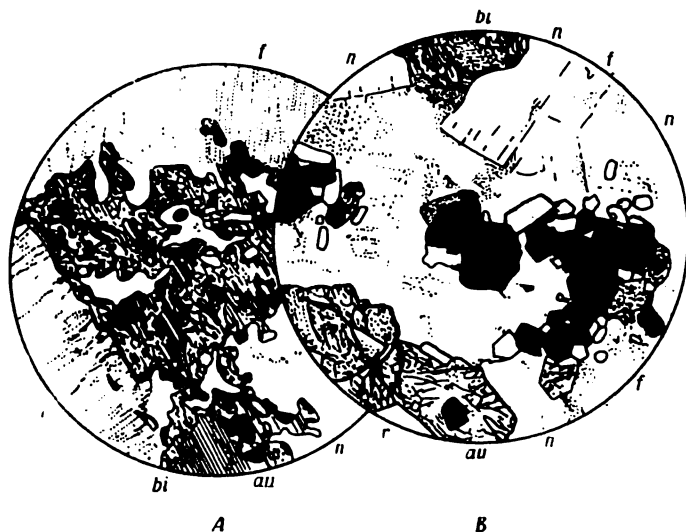


Fig. 84. Nepheline syenites, $\times 20$:

A—brown biotite (bi) and green aegirine-augite (au) in parallel intergrowth; perthitic feldspar (r) and pure nepheline (n) in the interstices; accessory apatite and magnetite. B—abundant apatite and magnetite, light augite and brown biotite, idiomorphic nepheline and cryptoperthitic feldspar

clase. Alkaline pyroxenes and amphiboles form the coloured constituents in most cases (Fig. 84).

Zircon and various zirconium and titanium silicates with more or less appreciable content of rare earths are frequently found in some nepheline and alkali syenites. Primary calcite and cancrinite are present in many nepheline syenites.

Rocks containing feldspar exclusively in the form of plagioclase (labradorite) are called theralites or nepheline gabbros (Fig. 85). Mica nepheline syenites are called miaskites. In miaskites and some other nepheline syenites a tendency to a banded gneissoid structure can be observed. In these rocks, the alkali feldspar is represented by orthoclase, microcline and albite separately or together, often in perthitic intergrowths. Nepheline forms colourless or slightly turbid hexagonal prisms; on decomposition, nepheline crystals give various soda zeolites and brightly polarizing prisms, fibres and aggregates of cancrinite.

Flakes and wedges of sodalite are often found with nepheline. It is colourless or very pale blue in thin section and is easily recognized by its isotropic character. Biotite is iron-rich, pale brown in colour and with intense pleochroism. The pyroxenes are pale green aegirine-augite and bright green pleochroic aegirine, sometimes as rims around augite; titanaugite occurs in the more basic varieties.

Amphiboles are represented by dark brown barkevikite, greenish blue arfvedsonite and the dark blueish-green variety, hastingsite. Iron garnet is found in some types of syenite. The usual accessory minerals of the nepheline syenites are sphene, iron ore minerals and apatite.

Leucite syenites are called *missourites*. The leucite is rarely preserved in a fresh condition, in the majority of cases being replaced by an aggregate of orthoclase and nepheline or zeolites.

Feldspar-free rocks with nepheline are also known: *ijolite* consists of 50 % of nepheline and the same amount of dark and ore minerals; *urtite* consists of 82-85 % of nepheline, 12-16 % of aegirine and about 2 % of apatite.

Most rocks of this family have a crystalline-granular texture. Some of them are characterized by a parallel disposition of platy feldspar crystals (miaskite).

As a rule, the ferromagnesian minerals have crystallized before the feldspars and the feldspathoid minerals later, but to this there are many exceptions. Moreover, intergrowths are very common, both among the coloured and among the light minerals, and sometimes even between the former and the latter. Some nepheline syenites again have a porphyritic structure from the development of large crystals of feldspar.

Nepheline syenites occur in the Kola Peninsula, in the Ilmen Hills, in the Minusin region, in the Ukraine and in the Caucasus. The coarse grained nepheline syenite of the Kola Peninsula is called *khibinite*.

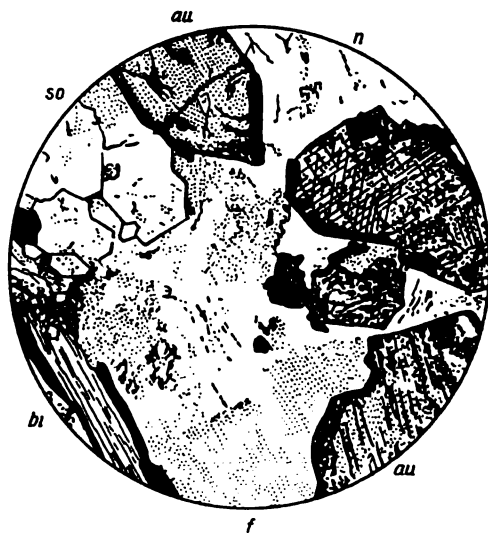


Fig. 85. Theralite, $\times 20$. Feldspar (*f*), nepheline (*n*), a group of sodalite crystals (*so*), augite (*au*) bordered with green aegirine and a little biotite (*bi*) pale in the centre and dark-brown along the edges

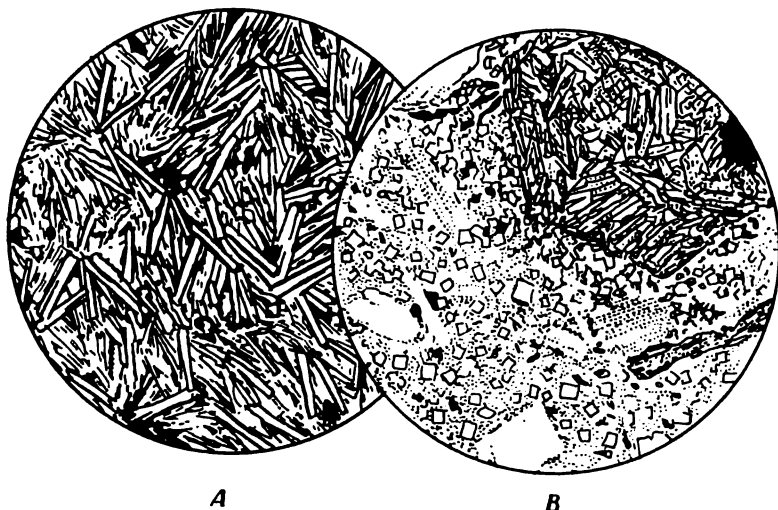


Fig. 86. Phonolites. $\times 18$:

A—trachytoidal phonolite. Consists mainly of sanidine in tabular crystals with aegirine in the interstices, and a few small crystals of nepheline; *B*—phonolite, rich in nepheline. A large phenocryst of hornblende has been altered by magmatic reactions into an aggregate of green aegirine

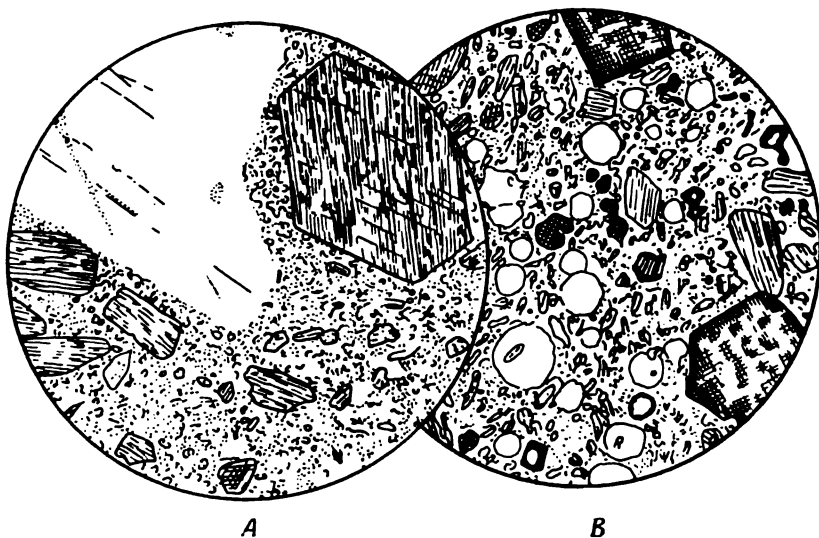


Fig. 87. Phonolite and leucitophyre, $\times 18$:

A—phonolite, phenocrysts of sanidine and aegirine in a groundmass of nepheline and sanidine; *B*—leucitophyre, phenocrysts of green aegirine bordered with a brown envelope of nosean and pure leucite, in a groundmass of aegirine-augite, nepheline and sanidine

Phonolites. The effusive equivalents of the nepheline syenites, independently of whether they are of Newer or Older Type, are called *phonolites* (Fig. 86). Alkaline effusive rocks, in which leucite and hauyne are found together with or replacing saundine and nepheline, are called *leucitophyres* (Fig. 87). They form flows or plugs.

A possible explanation of the origin of such rocks is based on the studies of the nepheline syenite massifs of the Khibine and Lavozero tundras, and ascribes then to the crystallization differentiation of a basic magma, or, more probably, to the enrichment of the magma by alkalis activated by volatile components, containing, first and foremost, water and fluorine. The abundance of volatiles is also confirmed by the presence of pegmatoid varieties and numerous pegmatite veins in the marginal areas of the massifs.

4. INTERMEDIATE ROCKS

Diorites

Diorites are deep-seated eruptive rocks containing a coloured mineral, most often hornblende, and intermediate plagioclase (from oligoclase to andesine), and this distinguishes them from gabbro, which is typified by labradorite or even bytownite and anorthite (Fig. 88).

Of a number of dark minerals, hornblende is the one most frequently present, often as idiomorphic grains; twinning is usual, and there is clearly seen prismatic cleavage (at an angle of about 125°).



Fig. 88. Diorites, $\times 20$:

A—augite diorite, idiomorphic crystals of light augite (au). Brown hornblende (h) is allotriomorphic and often has inclusions of augite and magnetite; B—diorite with idiomorphic hornblende; C—secondary growths of hornblende crystals in altered diorite. The feldspar is to a certain extent protected by calcite (ca) in which has grown a new generation of green hornblende

Dark brown biotite occurs in idiomorphic lamellae; they are sometimes intergrown with hornblende, are not usually rich in inclusions, and become green only when partially altered.

A parallel growth of augite and hornblende is sometimes seen in diorites; the crystal grain of augite forms the core, and around it accrues an envelope of brown hornblende, possibly secondary (Fig. 88*B*). Secondary hornblende can replace augite completely, and it is then difficult to distinguish it from primary.

The texture of diorites is hypidiomorphic-granular, granitic, sometimes porphyritic. The colour is grey, greenish-grey, sometimes dark grey, almost black. They are of frequent occurrence, usually as veins and do not form appreciable massifs.

Andesites

Andesites are the effusive equivalents of diorites (Fig. 89). They are the commonest lavas, and together with basalts, are almost the entire constituents of the volcanic fields of the Caucasus, Kamchatka and other regions. The colour of the rock is usually dark. Microlites of andesine are characteristic of andesites (in distinction to basalts which are typified by labradorite). The Older Type representatives of the andesites are called porphyrites (see Fig. 68). The dark minerals are augite, more rarely hornblende, usually present as phenocrysts and microlites.

Andesites are linked with dacites (andesite-dacites) and basalts (andesite-basalts) by gradual transitions. The texture of andesites



Fig. 89. Hornblende andesite, $\times 20$. Shows phenocrysts of zoned andesine and brown hornblende in a fine-grained groundmass

is known by the name hyalopilitic or andesitic, and under the microscope appears as a felted mass of acicular microlites of plagioclase permeated with glass; transitions to pilotaxitic and glass textures can be seen.

Alteration of andesites and porphyrites leads to the formation of epidote, calcite and chlorite, at the expense of the primary minerals, resulting in greenish or greyish alteration rocks (the greenstone porphyritic series of the Urals, Caucasus and Kazakhstan).

5. INTRUSIVE BASIC ROCKS

The typical intrusive basic rock of this group is gabbro, the constituents of which are basic plagioclase from anorthite to labradorite and some representative of the pyroxene or amphibole groups. The coloured constituents of gabbro are often represented by monoclinic pyroxene (Fig. 90). Bronzite and hypersthene gabbro are called *norite* (Fig. 91). Olivine often occurs along with pyroxene or amphibole (olivine gabbro and olivine norite). A constant accessory constituent (sometimes several per cent) is magnetite with a more or less appreciable content of titanium.

The labradorite in gabbro forms irregularly shaped plates with rather wide twin bands (albite twins), often intersected by fine pericline hatching. Orthoclase sometimes occurs (orthoclase gabbro). Weathered plagioclase can give sericite or kaolin, but more often

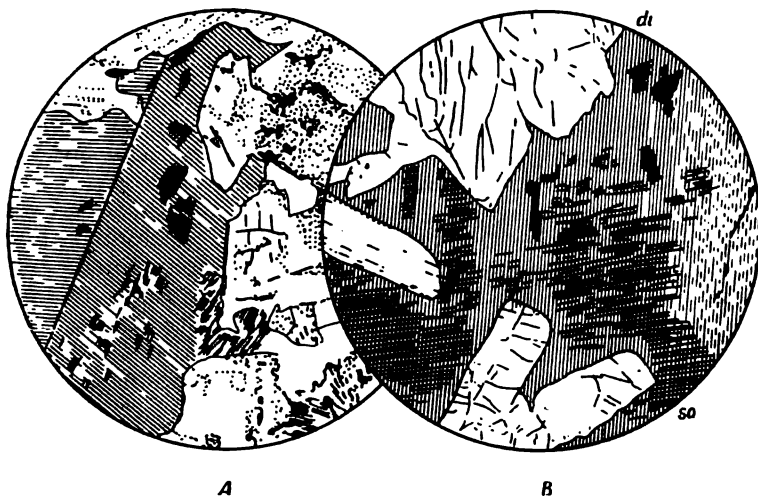


Fig. 90. Gabbro, $\times 20$:

A—quartz gabbro composed of augite, labradorite and slightly altered biotite and pure quartz. The augite is finely striated (diallage); B—the same, the augite in the field of view belongs to a single crystal

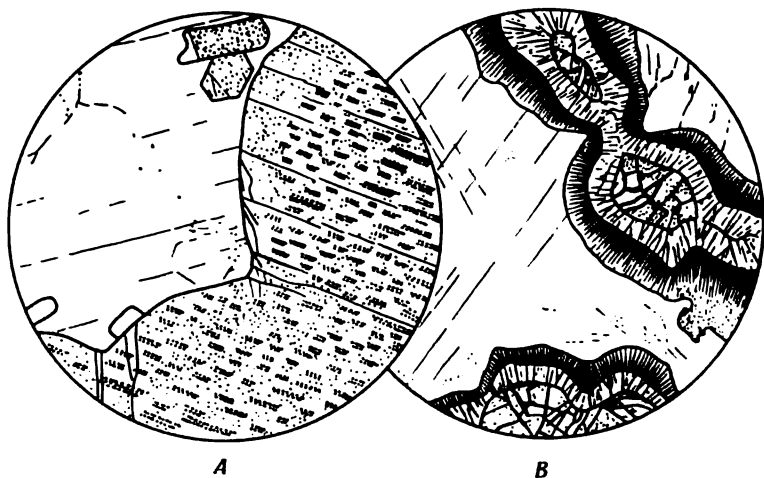


Fig. 91. Norites, $\times 20$:

A—the rock consists mainly of labradorite and hypersthene. At the top are crystals of apatite; B—olivine is surrounded by a double border consisting of enstatite (inside) and fibrous hornblende (outside)

it is converted into an aggregate of albite and zoisite, known as saussurite.

The pyroxene is most often augite, and forms irregular crystal plates or infillings between plagioclase crystals (ophitic texture). Decomposition of augite gives scales or fibrous aggregates of chlorite and serpentine. Another usual alteration is conversion to hornblende, which can be light green and fibrous (uralite) or dark brown and compact.

Primary hornblende is rarely found in these rocks. Dark brown hornblende is known in some norites. Brown or green biotite can be present as a primary or secondary mineral.

The orthorhombic pyroxenes, enstatite and hypersthene form somewhat rounded idiomorphic crystals. The alteration product of orthorhombic pyroxenes in gabbro is serpentine, forming the pseudomorphs to which the name bastite is given; it is pale green or yellowish with weak pleochroism and low polarization colours.

Olivine forms imperfect crystals or rounded grains, colourless in slice. They are often bordered with a rim of orthorhombic pyroxene where they are adjacent to feldspars (Fig. 91B). Serpentinization is a very characteristic alteration process. It begins along the edges and the interior cracks of a grain. Veinlets of granular magnetite separate out along the cracks during the first stage. The walls of the cracks are converted to pale greenish or yellowish serpentine, the serpentine fibres being disposed perpendicular to the walls, and giving straight extinction and low interference colours. The gaps are

occupied by residues of unaltered olivine, which can then be converted into another kind of serpentine, which is often a mass of tangled fibres and almost isotropic. The conversion of olivine into serpentine is accompanied by an increase in volume, causing the appearance of numerous radial cracks in the neighbouring minerals. These cracks are usually filled with isotropic serpentine.

A little quartz and orthoclase can be present in acid gabbros, sometimes in pegmatitic intergrowth (*quartz gabbro*). Accessories are ilmenite (with leucoxene as its decomposition product) and magnetite; apatite and sphene are also found.

The texture of gabbro is crystalline-granular. Equigranular gabbros are often distinguished by a great coarseness of grain and two types can be established: the first has its dark minerals more idiomorphic than plagioclase (gabbro texture); in the second, the feldspar has separated before the coloured constituents and is idiomorphic (ophitic or diabasic texture). Porphyritic gabbros are found sometimes.

An interesting texture arises in those cases when olivine separates in the early stages of crystallization of the rock; and it can subsequently partially dissolve in the more acid residual melt with the formation of orthorhombic pyroxene at its own expense, and which appears to be growing on to the remains of the olivine crystals. The pyroxene can sometimes be seen to react in its turn with the melt in the last stages of crystallization, and then a second rim is formed consisting of hornblende (Fig. 91B) (reaction rims). Similar borders can be formed in other cases as the result of a reaction between adjacent minerals even after the solidification of the rock.

Banding is found in some rocks of this family, bands relatively rich in feldspar alternating with bands enriched in dark minerals and with lenticles consisting mainly of dark minerals or especially rich ores of titanomagnetite. It has been suggested that this structure is the result of magma movements which took place during crystallization.

Many basic rocks can be regarded as varieties of gabbro. Thus *gabbro-diabase* butts against ophitic gabbro, being an ophitic basic rock with chlorite. A plagioclase-olivine gabbro bears the name of *troctolite*. *Teschenite* can be regarded as an analcime gabbro. Gabbros are linked with diorites by gradual transitions (with decrease in plagioclase number). An increase in alkali content leads to the appearance of *orthoclase gabbros*, and, when the amount of orthoclase becomes approximately equal to the amount of plagioclase, to the *monzonites*, which are already transitional to the group of the syenites and are sometimes called *gabbro-syenites*. An increase in the amount of dark minerals leads up to ultrabasic and ore gabbro. Titanomagnetite or iron, copper and nickel sulphides can be present as an ore mineral. Almost monominerallic plagioclase rocks bear the name of anorthosites, or labradorites (the labradorites of the Zhitomir district are excellent construction material, used for facing buildings).

Gabbro rocks are widespread in all regions where plutonic magmatic rocks occur, but they nowhere form such large massifs as the granites. Gabbro serves as a very good building material.

6. EFFUSIVE BASIC ROCKS

The foremost rocks of this group are the basalts and their Older Type analogues, the diabases, melaphyres and augite and labradorite porphyrites. Basalts are the commonest lavas of present-day volcanoes, as well as those of the Tertiary and Cretaceous periods.

Augite porphyrites and other Paleobasalts play a predominant role among the volcanic products of the Greenstone series of the Paleozoic and Precambrian.

In their external aspect, basalts are dense black masses, forming lavas in which plagioclase and coloured minerals are found in approximately equal quantities; magnetite or ilmenite is always present; when olivine is present in appreciable amounts (Fig. 92), the rock is called an olivine basalt.

Basalts occur as flows, thick sheets, plugs, sometimes veins. Flows and sheets consist of several superimposed layers of successive eruptions, and preserve a typical external slaggy zone.

Sheets may be separated by interbedded tuffs. The external zone of each sheet or flow is sometimes completely glassy. Columnar jointing is often excellently shown by basalts and has received the name of basaltic jointing.



Fig. 92. Olivine basalt, $\times 45$, without analyzer

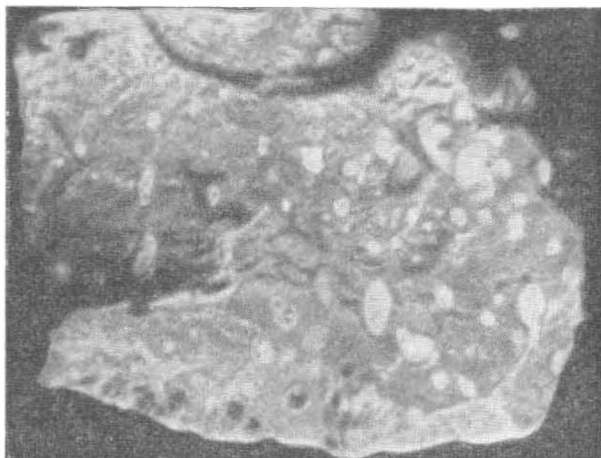


Fig. 93. Amygdaloidal greenstone

Because basic lavas are easily mobile and able to crystallize even in the absence of mineralizers and when the cooling is relatively rapid, basalts vary extremely in texture from holocrystalline and even comparatively coarse-grained *dolerites* to glassy basalt.

The commonest types of basalts are porphyritic rocks with an intersertal groundmass. The phenocrysts are olivine, sometimes plagioclase and augite. Basaltic glass (*tachylites* and *hyalomelanes*) is met with considerably less frequently than the acid glasses as a consequence of the ease with which a basic magma crystallizes.

Rocks formed by the submarine eruption of a lava, differ from basalt by containing a chlorite mineral which endows these Paleobasalts (diabases, pyroxene and labradorite porphyrites) with the green tint which has served as the basis for separating these rocks as the Greenstone group. In some cases, vesicles are present in these lavas and are filled with calcite, quartz and zeolites; in such a case, we have *diabase mandelstones* (Fig. 93). Diabase formations which have been especially thoroughly converted to greenstones are represented in the north-west shores of Lake Onega, in the Urals (Fig. 94) and in the Caucasus.

Spilites are extremely characteristic formations of ancient submarine volcanic series. They are soda-rich rocks in which the feldspar is represented by albite. Acid rocks of this type have been already described under the name of keratophyres.

Spilites are basic rocks with albite occurring both as microlites and phenocrysts. Entire horizons of such spilite-keratophyre lavas have been made out. Along with them occur characteristic submarine formations, pillow lavas. These are characterized by a spherical



Fig. 94. Diabases from the Urals

jointing, formed when the liquid lava welded out into the water. As a result of quenching, a glassy variety was formed in the external zones of these pillows, normal crystallization taking place only inside the pillows.

Ancient volcanic series are especially rich in various types of tuff.

7. ULTRABASIC ROCKS

Ultrabasic rocks are mainly found as intrusive formations. Effusive bodies of this composition are known in the U.S.S.R. only in the north of Siberia in very obscure geological conditions.

The characteristic feature common to ultrabasic rocks is their composition. They are made up only of dark minerals, and they are mostly almost monomineral pyroxene or olivine rocks.

Dunites consist almost exclusively of olivine, with the addition of chromite. Dunite is sometimes light green.

Peridotites are ultrabasic granular rocks, free from feldspars and feldspathoids, and consist of dark minerals, among which olivine predominates. The presence of an especially great amount of pyroxene produces *pyroxenites*. These rocks may be medium- or coarse-grained. Their colour is dark green, dark brown, almost black. Olivine is mostly converted to serpentine.

Of the pyroxenes, orthorhombic pyroxene is often present as well as diallage; characteristic accessories are titanium iron ore, magnetite, chromite and sometimes native platinum.

Serpentinites (ophiolites). When olivine and magnesium-rich pyroxene are decomposed, they are converted to serpentine (chrysotile or antigorite). The corresponding ultrabasic rocks rich in olivine

and pyroxene pass into chrysotile or antigorite rocks known as serpentinites.

The newly formed serpentine often preserves the external form of the decomposed mineral, making it possible to judge the original composition of the rock.

Some examples of such pseudomorphs are now given.

1. Serpentine produced from olivine, with "mesh" structure (Fig. 95A).

2. Coarsely lamellae serpentine after enstatite or bronzite, in recognizable pseudomorphs, and known as bastite (Fig. 95B).

3. Serpentine after non-aluminous hornblende, with reticulate structure.

This serpentine is finely fibrous, with fibres arranged perpendicular to the hornblende cleavage. The remaining part of the pseudomorph consists of serpentine which affects polarized light, and possibly made up of a felted aggregate.

4. Serpentine after pyroxene and having a scaly habit (antigorite serpentine).

On recrystallization, serpentinites give rise to talc, talc-chlorite, chlorite, talc-magnesite rocks; the very finely fibrous serpentine, chrysotile asbestos, is developed at their expense. A mass of magnesite is sometimes formed when serpentinites are decomposed.

Picrites and *picrite porphyrites* can be regarded as the hypabyssal members of this group, and are olivine-augite rocks containing an insignificant amount of chrome-diopside or other dark minerals sometimes a little basic plagioclase.

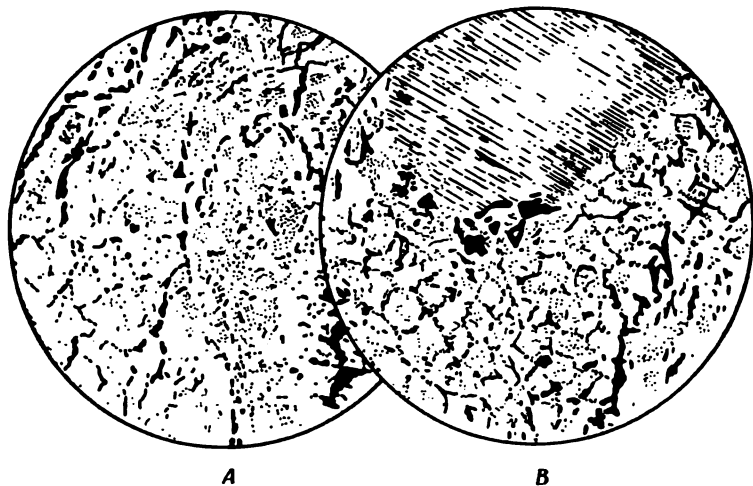


Fig. 95. Serpentine rocks, $\times 20$:

A—serpentinized dunite; B—altered peridotite. The big crystal is a bastite pseudomorph after enstatite; the groundmass shows mesh structure

A feature of picrites is their poikilitic texture, showing olivine grains scattered over the cleavage planes of the pyroxene or hornblende.

It is to this group that the diamond bearing *kimberlites* belong. These are breccia-like rocks which form the "explosion pipes" of South Africa and Yakutia in the U.S.S.R. "Explosion pipes" are small intrusive bodies, equidimensional in cross section, and passing into great depths; the kimberlite filling them consists of large crystals of serpentinized olivine together with grains of ilmenite and pyrope garnet, while phlogopite has sometimes been embedded in the basic essentially serpentine mass, with magnetite, apatite and picotite. Diamond is contained only in some pipes. Many pipes, although full of kimberlite, may be completely free of diamonds.

Industrial deposits of chromite, nickel and platinum are associated with ultrabasic rocks. Nickel silicates often fill in the cracks in weathered serpentines, forming nickel silicate ores. The primary source of platinum is dunite, more rarely pyroxenite.

8. OCCURRENCE OF MAGMATIC ROCKS IN THE U.S.S.R.

A knowledge of the occurrence of magmatic rocks in space and time is of practical as well as theoretical importance, especially in prospecting for and exploring deposits of useful minerals.

Acid intrusions played an exclusive role within the Soviet Union into the Archaean and Proterozoic eras. Special types of granite known as *rapakivi* and confined to the Precambrian are characterized by a considerable content of relatively large crystals of potash-soda feldspar, each framed by a rim of acid plagioclase (oligoclase). Many rapakivi massifs are found in regions where folded Precambrian rocks emerge at the surface. Examples are the Vyborg and Ladoga rapakivi bodies in the north-west of the European part of the U.S.S.R., the Korsun-Novomirgorod and Korosten massifs in the south.

An appreciable part of the magmatic bodies of Paleozoic age situated in regions of Caledonian and Variscan (Hercynian) folding proved to be basic rocks.

The following are examples: the Greenstone series of the Central Urals, consisting mainly of various effusives: porphyrites, diabases, trachyte porphyries; a relatively widespread development of ultrabasic and basic rocks in Kuznetsk Alatau; the Siberian traps (mainly dolerites, diabases, diabase porphyrites, basalts), which occur at intervals from the Upper Carboniferous to the Middle Jurassic.

The beginning of the Mesozoic era is also characterized by outpourings of basic magma and the injection of intrusive traps over a considerable area of the Siberian platform, without, however, excluding the injection of an acid magma as well.

In the Jurassic and Cretaceous periods various types of magmatic

activity have been established in a number of regions in the U.S.S.R. Thus, a thick volcanic series arose in Georgia during the Middle Jurassic, composed mainly of different kinds of porphyrites.

Lower Tertiary magmatic activity was manifested only in isolated regions of the U.S.S.R., but comparatively intensively in Transcaucasia and certain regions of the Far East.

Magmatism in the Neogene occurs in many places: the Trans-Carpathia, the Northern Caucasus, Georgia, Armenia, Azerbaydzhan, Western Turkmenia, Touve, Sayany, Southern Prebaikalia, Transbaikalia, the Far East.

Magmatic rocks of Quaternary age and varying in composition are seen mainly in the Caucasus and Far East. Present-day active volcanoes are concentrated on the western borders of the U.S.S.R.

9. THE PETROGRAPHY OF MAGMATIC ROCKS AND METALLOGENESIS

The present state of geological knowledge reveals a tremendous variety of controlling conditions — tectonic structures, types of rocks, metamorphism, metasomatism, the processes of oxidation and concentration of loose deposits and the ore formation associated with it.

There has recently been a refutation of the primitive ideas of some authors that there was a single granite type of ore-bearing magma, of a single type of ore-bearing intrusion in the form of batholiths, and of a single ore-bearing solution gradually ridding itself of its mineral burden as it travelled away from the batholith, and causing *ipso facto* a purely spatial zoning in the disposition of the deposits of the various metals.

It has been suggested that the separation of ore solutions in the magmatic melting pot is of a pulsating character, and the Soviet scientists S. Smirnov and Yu. Bilibin have played a large part in the development of this hypothesis. These solutions are now considered to enter the structures in which the ores are deposited, not all at once, but in portions at a time (depending on their accumulation in the place of melting), thus making possible repeated entries of new portions of solutions along the same channel. As the magma chamber congeals, the solutions systematically alter their composition mainly with time and only to a small degree in space. However, many problems of ore formation still remain unsolved.

Very important forms of mineral raw materials are associated with igneous rocks, and, consequently, the search for deposits and the evaluation of future industrial development of any region is impossible without an elucidation of the history of the development of a magmatic complex, or the determination of the conditions of deposition, the form and dimensions of the magmatic bodies, and the position of these bodies in the general geological structure.

Magmatism is always found in close association with the development of tectonic structures. The circumstances of formation of magmatic bodies—temperature, pressure, duration of the process, the material composition of the magma and enclosing rocks—determine the character of reaction taking place between the magmatic mass and its host rocks. They are also the primary causes of the separation of volatiles from the magma, and consequently of the concentration of the ore components whether contained in the magma itself, or as assimilation products of the enclosing rocks.

According to some investigators, the enclosing rocks represent a passive medium which has no effect on the course of ore formation, but participates in it only as a space which can hold a certain quantity of solutions depending only on its porosity or jointing. According to other authors, hydrothermal solutions do not carry any metals out of the magma, but become enriched in them by leaching them out of the beds of rock through which they flow. Thus, if the first ignore the role of the medium in which the ore formation process is taking place, the second group ignore the role of the magma. Both these points of view dominated the study of ore minerals until quite recently. It has now been established that deposits are created as a result of complex reactions between magma and the products of its reactions with the surrounding medium.

The absorption by a magma of fragments of its host rock undoubtedly affects the course of its differentiation. Assimilation sometimes leads to the formation of hybrid rocks or to the enrichment of the magma by additional material including ores. When the melts, solutions and magma sublimates enter in their turn into reaction with the surrounding rocks, they can enrich them with ore components, or form new ore minerals in the process of this reaction. The creation of commercial deposits of ore minerals is possible in both cases. Similar concentrations can also be created in the reservoir itself even in the course of its development, and without any reaction with the enclosing rocks.

The extension of our mineral resources is bound up with an understanding of the conditions of formation of ore deposits in any way associated with a magma, and with it, must be based on a thorough study of magmatic formations and detailed investigation of the results of their reactions with the enclosing rocks. The interaction of petrography and physical chemistry with the whole complex of ore sciences is especially important in the solution of this problem.

CHAPTER XII

Sedimentary rocks

When rocks are exposed at the earth's surface, they prove to be unstable and are subject to decomposition, or *weathering*.

Weathering takes place on a colossal scale.

Study of the geological structure of the crust shows that mighty mountain systems have repeatedly arisen on its surface, but weathering and denudation have annihilated them, and the surface of the earth's sphere has been converted into a gently undulating plain which perhaps existed only until the next stage of mountain building movements. The total amount of rocks which have been disintegrated, transported, deposited anew and subjected to some chemical conversion during the geological history of the earth comprises a series whose thickness is measured in tens of kilometres.

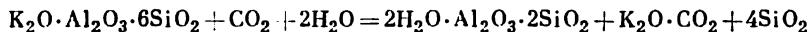
All of this tremendous amount of detrital, chemogenic and biogenic material deposited as sedimentary rocks consisted originally of magmatic rocks. The entire thickness of sedimentary rocks can be regarded as the final result of the redistribution of the constituents of magmatic rocks over the earth's surface.

Let us now discuss the more important of the processes by which sedimentary rocks are formed.

Weathering. Let us analyze the chemical weathering process in the case of the commonest magmatic rock, granite. Granite consists of feldspars, orthoclase and acid plagioclases, quartz, micas and accessory minerals.

In the oxidation zone, feldspars are converted first to sericite, and then to kaolinite, thereby freeing bases and silica which are removed from the point of reaction by ground waters.

This process can be represented schematically in the following form:



The weathering of mica proceeds in an exactly similar way, but here the amount of alkalies (and magnesium) will be a little less.

Table 11

Rate of Flow Required to Move Fragments of Various Size

Fragmental material	Mean diameter, mm	Rate of flow, m/sec	Fragmental material	Mean diameter, mm	Rate of flow, m/sec
Brick clay	—	0.08	Gravel	27.0	0.97
Fine sand	0.4	0.26	Gravel	54.0	1.62
Sand	0.5	0.28	Boulders	171.0	2.27
Sand	0.7	0.34	Boulders	323.0	3.25
Coarse sand	1.7	0.34	Boulders	409.0	4.87
Small gravel	3.2	0.46	Boulders	700-800	11.69
Small gravel	4.9	0.65			

the finest of fragments or dissolved substances can reach the central parts of these waters.

The ultimate effect of weathering and denudation is that fragments of quartz and other minerals resistant to weathering are concentrated in sands; dust-like particles form a sludge in suspension and comprise the argillaceous accumulations of marine waters. Substances which have been dissolved during weathering join the complex of salts in the oceans, and can be deposited from this aqueous solution, when the salt concentration reaches its critical value. Weakly soluble substances (as phosphoric salts, certain silicates, carbonates of lime) can reach their limiting concentration even in separate parts of the open sea, and precipitates of them can be seen in these places. It was originally thought that the main role in the formation of limestones, phosphorites and some other rocks could be ascribed to living organisms, the limestones being regarded as a simple accumulation of limy shells. Recent work, however (by N. Strakhov and his school), has shown that there are a great number of limestones which are chemical precipitates of limy material on the sea floor; besides, even the actual shells are completely dissolved if conditions are unfavourable for the accumulation of calcareous material in the place where they have been deposited.

At the present day, areas of the sea floor have been found where calcareous and dolomitic silt is being precipitated. The dolomitic silt is characteristically deposited in waters of high salinity.

Specific conditions are required for deposits of salts of high solubility. This takes place when the waters dry up completely, as for example in lagoons and lakes; for large salt deposits to be formed under these conditions, the loss by evaporation must be made up regularly with new portions of salt water.

The chemical composition of sedimentary rocks. During weathering, transport and deposition of weathering products, the material is subjected to thorough sorting in respect to composition, this involv-

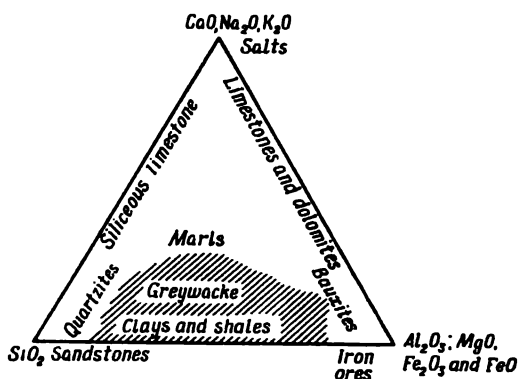


Fig. 97. Diagram of the chemical composition of sedimentary rocks. The composition of sedimentary rocks can occur anywhere in the triangle; the compositions of magmatic rocks occupy only the striated area

ing solubility, specific gravity or coarseness of grain. The result is the accumulation of new rock types very different from magmatic rocks, and differing to a considerably greater extent between themselves. The compositions of the most important rocks are given in a simplified form in the triangular diagram in Fig. 97. Sedimentary rocks occupy the whole field of the diagram, while magmatic rocks (in the shaded portion) are situated only in a small area near the base.

Diagenesis. In order to form a sedimentary rock, a sediment once accumulated has to pass through a stage of compaction and cementation. This process is bound up with the reaction of the sediment with the surrounding medium, or with the reactions of the constituents of the sediments between themselves. The following processes are involved: (a) solution and removal of soluble compounds; (b) deposition of newly formed minerals from the surrounding medium; (c) chemical reaction between the sediment and the surrounding medium; (d) dehydration or hydration; (e) cementation; (f) recrystallization; (g) ageing of colloids; (h) formation of concretions; (i) the development of more stable mineral assemblages at the expense of the less stable; (j) chemical reaction of the constituents of the sediment between themselves.

While they are accumulating, sedimentary rocks acquire their characteristic stratification, which distinguishes them from igneous rocks which are usually massive.

The stratification differs in form: inclined, undulating, horizontal, and so forth. An alteration in composition of the layers indicates an alteration in the conditions of accumulation of the deposit (a slowing up or acceleration of the current, an alteration in the depth of water, an alteration in composition of the constituent material).

The presence of detrital material, rolled pebbles and mineral grains, as well as the occurrence of various fossil remains or petrified plants

and living organisms is an assurance that we are dealing with a sedimentary rock. The actual composition of the rock is also often characteristic, especially when the minerals occurring are formed only under the specific conditions prevailing at the earth's surface, as for example, salts, gypsum, glauconite and similar substances.

Sedimentary rocks are themselves often commercially valuable (iron and manganese ores, phosphorites, clays, many building stones, and so forth) or else take a part in the segregation of such materials.

The total value of the commercial materials obtained from sedimentary beds is many times in excess of the value of those obtained from all other sources. For example, all mineral fuels (coal, petroleum, and natural gases) and the greater part of industrial aluminium ores are associated with sedimentary rocks; practically all ores of manganese, and a considerable portion of iron and nickel ores, as well as the overwhelming mass of mineral fertilizers are of sedimentary origin. Half of the world's output of gold, platinum and tin is obtained from unconsolidated sediments (placers). Building stones, refractories, acid-resistant material, ceramic, glass and optical raw materials, the various types of chemical raw materials, ballast and road-making materials mostly occur among sedimentary rocks and are worked from them.

Hence the importance of the science of sedimentary rocks in the study of the raw materials of industry. The petrography of sedimentary rocks plays no less a role in resolving the problems of hydrogeology, water supply, engineering geology and in paleogeographic investigations. With it, beds of ancient sediments can be shown to be the same or different, which is very important when studying non-fossiliferous beds, and also in regions where the geological structure is complicated.

The alteration with time of the type of detrital material entering a given region of sediment accumulation is a basic factor in micro-petrographic correlation. As a result, each bed or group of beds has its petrographical peculiarities. This is explained by the fact that weathering and denudation start immediately on all new beds of rocks, and implicate them in the process of sediment accumulation. The most rapid changes in the petrographic composition of accumulating sediments take place in times of mountain building.

1. METHODS OF STUDY OF SEDIMENTARY ROCKS

The great diversity in mineral composition of sedimentary rocks makes it necessary to study them by different methods, but in all cases, microscopic investigation in thin section takes first place.

For carbonate rocks (and sometimes for salts and calcareous sandstones) the method used is to dissolve the predominantly carbonate

portion in weak acid (in the case of salts, solution in water) and to study the insoluble residue under the microscope.

When studying sands and sandstones, their constituent grains are separated according to their specific gravity (in heavy liquids, for example) using electromagnetism and other physical methods, and the different fractions are studied under the microscope.

Coal (and sometimes limestones) are best studied in polished sections after etching.

Clays are good subjects for thermal analysis, and have latterly been investigated by electron microscopy.

The granulometric composition is widely studied by the use of sieves and by measuring the settling rate of the grains in water. The immersion method plays an essential part in studying the petrography of sedimentary rocks.

Sedimentary rocks thus require a wide variety in methods of investigation depending on the composition of the rock, and the objects aimed at in each particular case.

2. CLASSIFICATION OF SEDIMENTARY ROCKS

A generally accepted detailed classification of sedimentary rocks does not at present exist. However, a distinction is always made between mechanical deposits, chemical precipitates and rocks of organic origin. The difference of opinion concerning the individual schemes lies in the interpretation of the scope of any one group and whether a given rock should be assigned to this group. Numerous transitions exist between these groups. Let us imagine, for example, a sandstone where the cement consists of calcite, while the sand grains are quartz. It is not so simple to decide whether this is a chemical precipitate or a detrital rock. It is difficult to indicate a precise boundary between a sandy limestone and a calcareous sandstone. Such a boundary could only be arbitrary.

A group of sedimentary rocks of volcanic origin stand by themselves. This group consists of volcanic tuffs formed directly from volcanic eruptions, and various tufaceous sandstones and tuff breccias, which are the products of cementation of loose material of volcanic origin.

3. DESCRIPTION OF SEDIMENTARY ROCKS

Detrital Rocks

Various unconsolidated or cemented types consisting of fragments of rocks and minerals are assigned to this group. These rocks are further subdivided according to grain size and the composition of their cement.

Coarsely fragmental rocks, or *psephites* (the fragments are coarser than 1-2 mm): gravel consists of rolled fragments from 1 to 10 mm in size; shingle, or coarse gravel, of rolled fragments from 10 to 100 mm in size; rubble, of angular loose fragments; boulders are large rolled fragments; conglomerates are cemented aggregates of coarse rolled fragments—boulders, shingle, gravel; breccias are cemented angular fragments, or rubble.

Medium fragmentary rocks, or *psammites* have a grain size between 0.1 and 1-2 mm. Sands and sandstones belong here. Sands are unconsolidated aggregates of mineral grains of diameter not more than 1-2 mm. They are distinguished by origin: marine, lacustrine, fluvial and continental (desert); by grain size: coarse-, medium- (from 0.1 to 1 mm) and fine-grained; by composition: quartzitic, quartz-feldspathic (arkose), glauconitic, etc.

Sandstones are cemented sands. The cement can be very diverse, for example, calcareous, argillaceous, quartzitic, gypsum, marly, glauconitic, limonitic, bituminous.

According to the composition of the fragments, sandstones, and also sands, can be monomineralic, biminerale (arkose sandstones have fragments of quartz and feldspar) or polyminerale (polymict). Greywackes can serve as an instance of the latter: they are polymict psammitolites, formed by the disintegration of basic magmatic rocks and rocks related to them in mineral content. They are often of a grey colour, hence their name.

Wackes are distinguished as diabase, basalt, melaphyre and so forth, according to the name of the rock from which they originated.

Finely fragmented rocks, or *siltstones* are rocks consisting of grains 0.10-0.01 mm in size. Here belong loess and tremendous masses of variegated sandy-argillaceous rocks. In loess, fine fragments of quartz are nearly always present, as well as orthoclase, plagioclase, muscovite, biotite, magnetite, amphibole, pyroxene and other minerals. Calcite is always present, either as finely dispersed particles, or as thin envelopes around the little grains of quartz and other minerals, frequently in the form of calcareous concretions or loess dolls.

The most finely fragmented rocks, or *pelitites* (lutites) are fragmentary rocks, consisting of grains smaller than 0.01 mm. This is the commonest group of sedimentary rocks, forming about 4% (by weight) of the whole crust. They are usually termed clays. Due to the especially small particle size, the mineral composition of pelites is investigated not only by mechanical, chemical and microscopical methods, but also by thermal analysis (the plotting of curves of heating and dehydration), X-rays, and the electron microscope.

In the dry state, clays have an earthy character, and sometimes considerable strength. Mixing with water induces plasticity, and after firing they become stony, hence the use of clays in the ceramic industry.

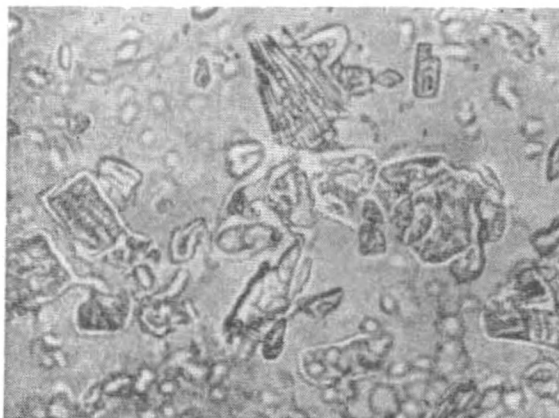


Fig. 98. General appearance of kaolin grains. Fine fraction (less than 0.01 mm); $\times 365$. An immersion preparation (liquid, $N = 1.542$). Vermiculite intergrowths and individual lamellae of kaolinite are shown

The basic constituents of clays are usually specific mica-like silicate minerals, grouped together as the "clay minerals". The most important among them are kaolinite, hydromicas, montmorillonite, various ferromagnesian clay minerals, allophane, pyrophyllite, and so on. The mineral composition of clay controls its properties and practical use.

Minerals of the kaolinite group (Fig. 98) and the hydromicas are the basic ingredients of the kaolin clays, the main raw materials for the production of refractories and fine ceramics. These clays are also used as fillers, for instance, in the production of rubber, paper and pencils.

The montmorillonite minerals form characteristic floridine and bentonite clays, used as whitening agents (Fuller's earth), various absorbers, plasticisers, etc.

Ferruginous clays, if not nickel or iron ores, are used as building material or for pottery.

The magnesium clays, pyrophyllite and allophane are rare, and have practically no independent importance. They are components of clays made up of other minerals, and impart special features to a deposit.

Almost all clays must be regarded as polyminerale rocks. Besides the predominant clay mineral characteristic of a particular type of clay, numerous other minerals are always present (quartz, muscovite, feldspars, etc.) and have an indisputable effect on the properties of the clay and sometimes determine how it shall be used.

The genesis of clays is known only in the most general outline. One of the causes of this is the inadequate knowledge of the mineral

composition of clays, their diversity, and the variety of conditions under which the same minerals can form.

Two genetic types of clays can be distinguished:

(1) eluvial clays, which have been formed by the weathering of some sedimentary or igneous rock; typical pseudomorphs after the rock are seen;

(2) sedimentary clays, formed as a deposit on underwater bottoms, and found among other sedimentary stratified formations.

Large deposits of refractory clays are known in the Donbass (Chasov-Yar), in the Urals, and some other regions of the U.S.S.R. Kaolin is found in the Vinnitsa and Dnepropetrovsk regions and elsewhere.

The decolourizing montmorillonite clays possess the ability to absorb organic material of high molecular weight. On this is based their application in oil refining, and the cleaning of vegetable oils and wool (Fuller's earth). Deposits of such pure clays are known in Georgia and Turkmenia. Clays with a large content of iron oxides are extraordinarily widespread. Low-melting brick clays and various ochres belong to this group.

Clays mixed with sand are called loams; loams of glacial origin crowded with boulders are called boulder clay. They are very characteristic of glacial deposits and are widely developed in the central areas of the U.S.S.R. (in the Moscow region and in the Kalinin and Tula districts).

Bauxite is also a pelitic rock, and represents an aggregation of various hydrated oxides of aluminium, formed in small lakes or shallow seas. Bauxite is the main source of aluminium.

Marl is a mixture of clay and limestone. The texture of marls is sometimes chalky, sometimes compact; it sometimes acquires a flaggy jointing. It effervesces with HCl, leaving a patch of clay. Is an excellent material for the preparation of portland cement; such marls are found in the neighbourhood of Novorossiysk.

Compact cemented clays are known as argillites. Apart from the clastic minerals and silica, argillites often include sericite, chlorite and carbonates.

Argillaceous rocks with slaty structure are called shales and slates.

Chemical Precipitates

Chemical deposits are formed by the precipitation of mineral particles from solution, and are preserved only when they are protected from the solvent action of water, as, for example, under an argillaceous cover. Haloids, sulphates, some carbonates, siliceous material and some brown iron ores can separate from solution.

Sodium chloride is the main constituent of the salts dissolved in the water of the seas, some lakes and springs; when these waters or parts of them dry up, sodium chloride separates out as deposits of

halite. Rock salt is usually colourless, but it often contains different impurities which give it various colours. Small amounts of Mg, $\text{Ca}(\text{CO}_3)_2$ and CaSO_4 are always present. Intercalations of halite with gypsum and anhydrite are often recorded, and sometimes particles of argillaceous material are mixed with the salt. In the upper horizons of such deposits, the easily soluble salts as carnallite, sylvinite, kieserite are often present. In the U.S.S.R., rock salt deposits are found at Artemovsk (in the Ukraine), in the Iletskaya Zashchita (the Orenburg district). The largest deposits of potash salts in the world are found in the Solikamsk region.

Gypsum is often formed at the same time as rock salt, and is sometimes found occurring independently in Permian and Tertiary beds as a rule. The gypsum is fine-grained, compact and fibrous in texture. As has been mentioned, some present-day limestones are chemical precipitates.

Phosphorite beds, worked for phosphate fertilizers necessary for agriculture, must be regarded as chemical deposits.

According to a hypothesis of A. Kazakov, phosphorites are formed in a definite phosphorite facies in a littoral marine zone, in the presence of currents bringing dissolved phosphates from the oceanic depths, and the phosphates are precipitated as an effect of a decrease in the partial pressure of carbonic acid.

The sedimentary iron ores are the cause of much dissension. Ya. Samoilov regarded them as bacterial deposits, an accumulation of the products of activity of iron bacteria; but recently N. Strakhov has shown that at least some, and maybe all deposits of sedimentary iron ores, as well as sedimentary manganese ores and bauxites, can be considered chemical precipitates, unconnected with bacterial life.

There is no doubt that aggregations of various silicates such as glauconite, and ferruginous chlorites and zeolites are chemical precipitates. Ferruginous chlorites sometimes accumulate in such quantities (thuringite) that they are used as iron ores (Alsace).

Glauconite proves to be a highly interesting mineral. It used to serve as a cheap and unstable green colour, but it has now become important in the economy as a natural water softener (the raw material of permutite).

Glauconite is a marine chemical precipitate and belongs to the group of the hydrous micas. Its composition is very variable, from the dark-green ferruginous variety to almost colourless and poor in iron, the ferruginous dark-green glauconites being formed in shallow water conditions, and the near colourless variety in the deepest parts of the ocean. In this range of glauconites, the successive variation of both chemical composition and colour is accompanied by changes in their technological properties, in particular, their capacity for kation exchange.

Certain rocks of the Moscow basin, previously accepted as ordinary clays, have now been shown to be actually wholly made up of

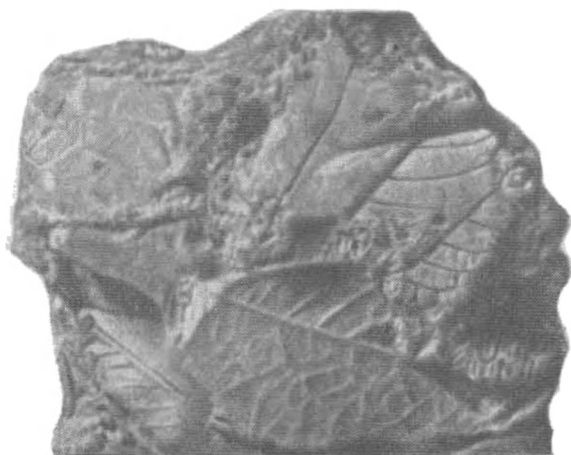


Fig. 99. Calcareous tufa (travertine) with leaf impressions

finely dispersed precipitated glauconite. This type of glauconite occurrence was not known at all previously.

Calcareous tufa and siliceous sinter are examples of peculiar products of chemical precipitation.

Calcareous tufas are formed when carbonic acid is given off from the water of the springs containing it. This lowers the solubility of the carbonic acid, and calcite is precipitated from the water in the form of calcareous tufa, or travertine. This is a porous, vesicular incrustation often used as a decorative material (Fig. 99).

Siliceous tuffs or geyserites belong to the *siliceous sunters*, and are deposits of amorphous silica from hot springs, mostly from geysers.

Sedimentary Rocks of Organic Origin

Remains of living organisms play a great part in the composition of many sedimentary rocks. When they participate in sediment formation, they are one of the important factors ensuring the trend and irreversibility in the formation of a sedimentary rock.

Organisms change the composition of the atmosphere, and also the gaseous processes in water masses; both during their life and after their death, they determine to a considerable extent the oxidation-reduction potential of the sediment forming medium and its characteristic pH value. As a result of this, weathering and sedimentary differentiation turn out not to be purely physico-chemical processes, but to some extent, biological processes.

The shells or skeletons of rock-forming organisms fall into three groups based on chemical composition: siliceous, carbonate and phosphatic.

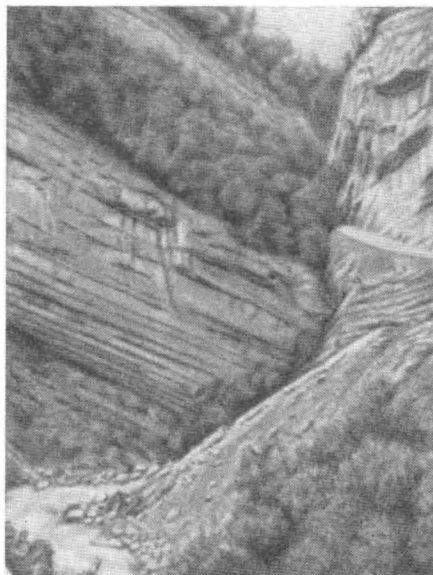


Fig. 100. Jurassic limestones in the gorge of the river Uruk (the Northern Caucasus)

The carbonate rocks play a great part in the composition of the crust, being represented by limestones (Fig. 100), dolomitized limestones and dolomites.

According to texture, limestones are distinguished as crystalline (marbles), compact, earthy, chalky. Very fine homogeneous pelitomorphic limestones form lithographic limestones *. The earthy variety of limestone, known as chalk, consists of the shells of very simple animals (rhizopods from 0.008 to 0.1 mm in diameter), of the spherical bodies of algae, known as coccoliths, and minute crystals of various minerals (quartz, zircon, feldspar).

If the clay content of a limestone is not more than 15%, the rock is called a marly limestone. If there is up to 30% of clay, it is a marl; and if more than 30%, an argillaceous marl.

Dolomitized limestones and dolomites (Fig. 101) contain dolomite as well as calcite. Dolomite formation is mostly associated with a very late displacement of calcium by magnesium. Sometimes, however, but evidently very rarely, they are also formed by the direct deposition of precipitates containing a significant amount of $MgCO_3$.

* Homogeneous fine-grained limestones, which are usually precipitate, are used exclusively as lithographic stone in polygraphy. In the U.S.S.R. their deposits are known in the Caucasus. Fine-grained limestones are called "lithographic limestones".

Carbonate rocks comprise more than half the volume of all sedimentary formations of the Russian platform. In Paleozoic deposits they are represented predominantly by limestones and dolomites, whose content reaches 60% of the total volume of deposits.

In Mesozoic and Cenozoic deposits as a whole with a volume considerably less than that of the Paleozoic, the commonest carbonate rocks are chalk and marls; there are few limestones and dolomites. The content of carbonates is here comparatively small and does not exceed 20% of the total volume of deposits.

The distribution in time of carbonate and terrigenous rocks on the Russian platform is not fortuitous, and reveals systematic features and periodical repetition. In the course of each of three successive tectonic stages (the Caledonian, Hercynian and Alpine), there is consecutive change of lithological complexes (Fig. 102): at the beginning of the tectonic stage deposits of clastic rocks predominate; in the middle, the dominance passes to the carbonate formations; and at the end, the dominating role is again acquired by the clastic rocks.

Parallel to this systematic change in the quantitative relations of the different types of sedimentary rocks in the course of geological time, each stage is characterized by its own particular composition of deposits, and consequently by its own geochemical processes.

The changes taking place in the composition of the carbonates did not bear the character of a simple repetition. Thus, the content of magnesium in carbonate rocks obviously declines from the Prote-

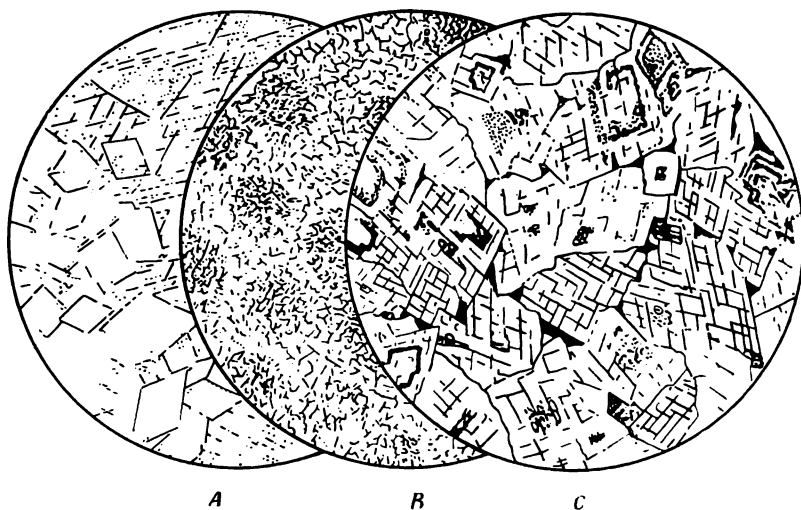


Fig. 101. Dolomite rocks, $\times 20$:

A—dolomite rhombs in a calcite groundmass; B—dolomitized limestone; C—dolomitized limestones with ferruginous material marking the growth zones in the dolomite crystals

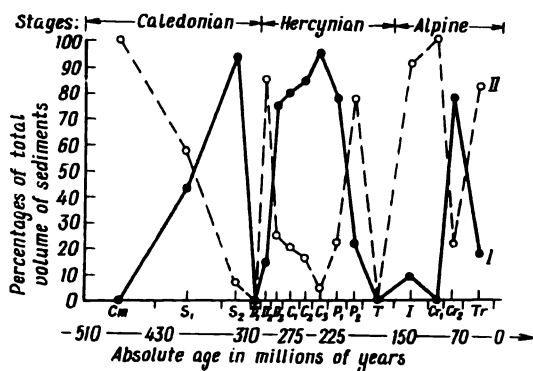


Fig. 102. Change in the percentage content of carbonate *I* and detrital *II* rocks in the course of the Caledonian, Hercynian and Alpine stages of sedimentary accumulation in the Russian platform

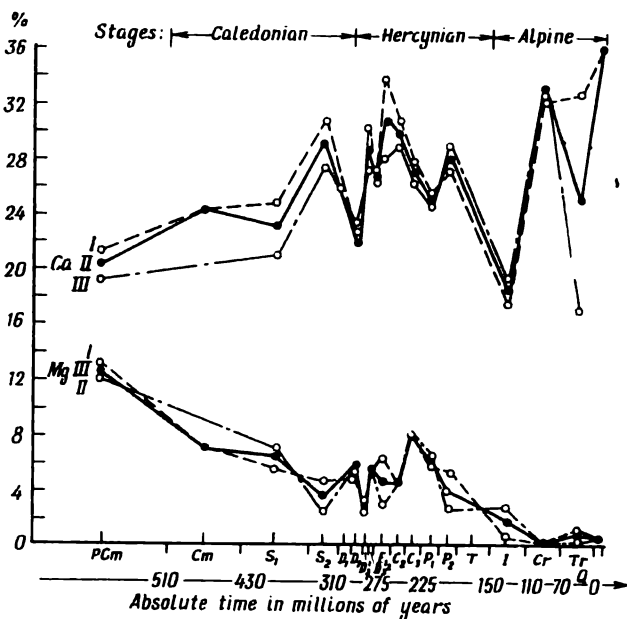


Fig. 103. Change with time of the average Ca and Mg content in carbonate rocks of the Russian platform:

I—curves constructed from averaged published analyses (7636 analyses for Ca and 7396 for Mg); *II*—curves constructed for the mean values of adjusted average samples (97 analyses of adjusted average samples from 3659 specimens); *III*—curves constructed by averaging all analyses (published and adjusted average samples)

zoic, and in the carbonates of Tertiary age, the magnesium content is on an average 15 times smaller than in the Precambrian (Fig. 103).

Evidently, the evolution of carbonate composition revealed in the case of the Russian platform also took place in some form or other in all the other carbonate deposits of the ancient seas of the earth.

Tripoli earth, or diatomite, is a light grey or yellowish rock, soft to the touch and easily pulverized. It consists of the siliceous frustules of diatoms; compacted and finely foliated diatomites are called polishing slates. Tripoli earth occurs predominantly in deposits of the Tertiary system in the U.S.S.R., in Povolzhye, in the Moscow region, in Zakavkazye.

Siliceous gaizes consist of hydrated silica (up to 97%). The silica of gaizes represents an accumulation of diatom frustules, radiolarian shells, the siliceous sponge spicules, and secondary structureless silica obtained from the alteration of organically formed SiO_2 .

Coal, combustible shale and petroleum also belong to the sedimentary rocks. Chapter XIII is devoted to the most important facts of coal petrography.

Volcanic Tuffs

Volcanic sedimentary rocks form a special group of clastic sedimentary rocks (Figs. 104 and 105). They are formed by the accumulation and cementation of fragments of various volcanic materials—ashes, sand, bombs. They occur interbedded with sedimentary rocks. If these fragments accumulate from submarine eruptions along with normal fragmentary material and the remains of organisms, tufogeneous sedimentary rocks are formed. The amount of volcanic sedimentary rocks comprising older effusive deposits is usually much greater than the amount of the actual effusive material.

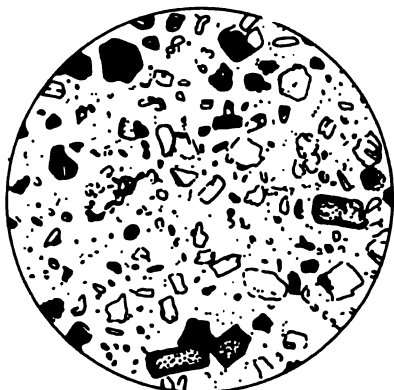


Fig. 104. Volcanic ash, $\times 20$. Crystals of plagioclase, magnetite, hypersthene and augite; a little glass



Fig. 105. Volcanic tuff, $\times 50$. The groundmass has fluidal texture. The inclusions are rock fragments and feldspar crystals

CHAPTER XIII

The petrography of coal

The vegetable origin of coal was pointed out long ago by M. Lomonosov (*Concerning the Layers of the Earth*, 1763).

The material from which coal is created accumulated either in marshes from the remains of terrestrial plants (humic coals), or in lakes from algae and the remains of animal plankton (sapropelic coal), or in lagoons from shallow water organic muds (combustible shales).

Peat formation begins in the deep horizons of a peat bog, where there is practically no access of atmospheric oxygen; it is a decomposition process in which oxygen takes no part, the products being gradually enriched in carbon. The lignin-cellulose tissues of the higher plants are converted into humic acids, which colour bog water brown. The process is called *humification*, after its main product.

The decomposition of the remains of lower plants, which consist mainly of proteins and fats, takes place with the aid of anaerobic bacteria, and leads to the formation of a jelly-like mass or sapropel. The proteins and fats are converted to fatty acids and bitumin, and the process is *bituminization*.

Three main periods can be distinguished in the formation of coal.

In the *first period*, known as the peaty period (before the bed is covered), the plant remains are decomposed and altered, mainly by biochemical processes taking place in the vegetable mass with the help of anaerobic bacteria.

In the *second period* (subsequent to the covering of the peaty layer with a roof), there takes place a physico-chemical alteration of the plant substance during the diagenesis process. These changes are directed towards an increase in carbon content, a lowering of the oxygen content, dehydration and compaction of the peat and its conversion to brown coal.

In the third period, the brown coals are converted into coal and anthracite as the result of metamorphism.

Mineral substances which chance to be in the peat bog as extraneous matter and solutions act as catalysts, or react chemically with the organic material, and their presence is in some way reflected in the properties of the coal.

Mineral coal is thus, first and foremost, a sedimentary rock consisting of an organic part (the coal substance) with added minerals; the composition and properties of a coal are controlled by the original material, the conditions under which it accumulated, and the method by which it was converted into coal.

Four petrographic constituents are distinguished in mineral coal: fusain, durain, clarain and vitrain (Fig. 106). They are not at all equivalent in their genetic and practical relationships. Fusain and vitrain usually appear as lenses and inclusions of limited size. They both appear to be homogeneous substances. Fusain retains a distinct plant structure, while vitrain either preserves no such structure at all, or preserves it in a concealed form, which can only be made out at great magnification. Durain and clarain, on the other hand, are complex aggregates, consisting of a groundmass and the essential constituents (preserved portions of plants) in any proportion. Fusain and vitrain in the form of more or less fine remnants can also be ingredients.

Elements of clarain can enter into the composition of durain, just as portions of durain can be found in clarain, but neither occurs

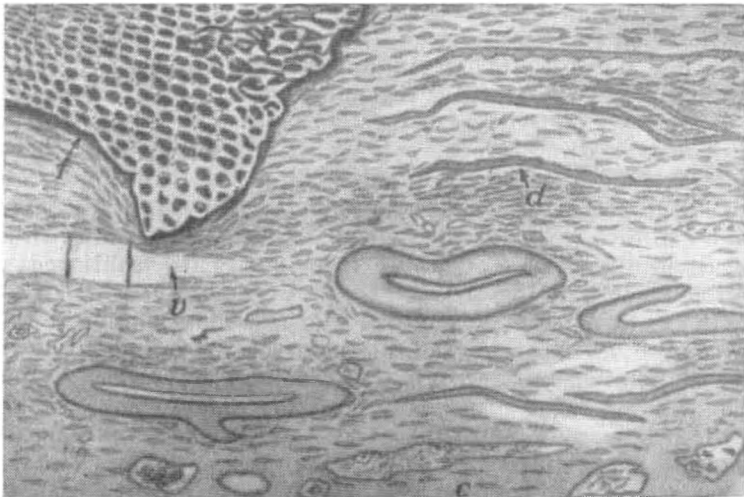


Fig. 106. The petrographic ingredients of coal:

f—fusain; d—durain; c—clarain; v—vitrain



Fig. 107. Fusain in reflected light, $\times 90$



Fig. 108. The microstructure of durain in reflected light. $\times 90$

in vitrain and fusain. There is thus the same relationship between these two pairs of coal constituents as between simple and complex rocks.

Fusain is reminiscent of charcoal macroscopically, and belongs to the dull varieties of coal. A transparent slice under the microscope usually reveals coherent cell structures. The cell walls are black and opaque, while the cell openings remain free and transparent or with an infilling of the mineral mass. In reflected light (i.e., on the polished surface of the coal) tissue structure is also seen, but the position is now reversed: the cell walls are bright white and slightly yellowish, while the apertures are black or dark grey (Fig. 107).

Sometimes swellings of the cell wall are seen, so that the apertures become smaller, and in places completely disappear. Sometimes the cell tissue is destroyed, as if shattered by a lateral pressure, especially in thin-walled cells (stellate or arcuate structure). In many cases, all gradations of structure from entire to shattered occur in one thin section.

Fusain is most frequently formed by the woody portions of sporophytes and gymnosperms. Fusain consisting of fragments of angiosperms is found in young coals. But even in the largest masses of fusain an entire organ of a plant is almost never seen. The plants appear to have suffered intense comminution before deposition.

The second peculiarity of fusain is that small clots and lenses of it are usually disseminated throughout the entire extent and thickness of a bed, indicating sorting action while the bed was being formed. However, aggregations consisting entirely of fusain can be observed in some coal seams.

Durain is the second dull ingredient of coal. It has a granular texture, and under the microscope always reveals a non-uniform constitution. It consists of a structureless groundmass, with various inclusions embedded in it; these are plant remains (spores, cuticle, resin bodies, remnants of plant tissues), referred to as the essential constituents (Fig. 108). There is less groundmass than essential constituents.

The durains most typical of Carboniferous coal contain a preponderance of little capsules, which are spore exines (their contents have been destroyed). Mesozoic and Tertiary coals often contain pollen and cuticle, spores rarely. The other constituents are seldom found.

Cuticle is preserved rather frequently, since it consists of cutine, an extremely stable substance. It is the external cutinized layer of the epidermis of leaves, branches and young shoots. Cuticle is impermeable to water, and protects the plants from excessive evaporation. The cuticle layer is interrupted only above the stomata (apertures in the epidermis for breathing, mostly on the lower side of the leaf).

In transmitted light, cuticle appears as a transparent yellow band in Δ and Γ grade coal, and orange or almost red in \mathcal{H} grade coal; it becomes invisible in coking and hard coals. In vertical section it has the appearance of long elongated lines or closed figures; in horizontal section it may seem to have a cellular structure with or without stomata. Cuticle is not divided into cells, what it shows is only the impression of cells on the upper layer of the epidermis (because of dissimilarity in the thickness of the cuticle layer above the cells and the spaces between them). These impressions have straight walls in Paleozoic plants, and undulating ones from the Mesozoic onwards. In reflected light, the cuticle has slight relief and is distinguished by shades of dark grey.

Since the cuticle is similar to the exines of macrospores in thin section (yellow remnants), it is important to be able to distinguish between them by referring to the following diagnostic features of cuticle: a serrated internal edge; the ratio of the thickness of cuticle to its length (1 : 50) is less than the ratio of the width of the exine of an entire macrospore to its length (1 : 20, more often 1 : 10); the thickness of cuticle is not the same in various places; the terminations are angular (spores have rounded forms).

Resin bodies. These are yellow in colour, transparent in thin section, usually oval formations, possessing no structure and in this distinguishable from spores and algal. They are often found in both durains and clarains. In transmitted light, these resin bodies are lemon-yellow or orange-yellow, structureless, oval, sometimes polygonal formations. In reflected light, they are grey and a shade darker than the groundmass. They are distinguished by their smoothness, distinct outlines and clear relief.

Woody tissues can enter into the composition of durain as microscopic fragments, as fusain, or as a brown or red-brown substance which is transparent in thin section and preserves traces of cell structure. In the latter case, several successive stages can be noted. First, the cell contents vanish, and the cell wall suffers distention and is converted into a gelatinous colloidal substance; cellular tissues with hollow apertures result; this structure has received the name *xylain*. Second, as a result of further distention of the wall, the cell almost merges into the general mass, leaving only small and irregular apertures in places. It can be guessed rather than seen from these apertures and the colour as a whole, and sometimes only from the general lenticular form, that a given fragment originated in a scrap of tissue. At this stage of gelatinization, the tissue is termed *xylovitrain*. Finally, if even these apertures are filled in, and gelatinization sets in over the entire fragment, the result is *vitrain*. Differences in shade sometimes show up the pattern of cell walls very clearly; in transmitted light, the walls are darker than the humus-gel infillings of their internal cavities. Such vitrain is called structural vitrain; in the opposite case, i.e., when the cell structure is undiscernible, it is called structureless vitrain, or simply vitrain.

Plant tissues can thus be found in coal in the form of fusain, xylain, xylovitrain or vitrain, and can be structural or structureless. Therefore, as well as the durains consisting purely of spores and cuticle, there can also be dull coals consisting predominantly of fusain, xylain or some other vegetable elements.

Some investigators used to consider the opacity of the groundmass as the distinguishing feature of durain. But a transparent groundmass has now also been shown to be to some extent a possible participant in the structure of dull coals. Thus, either form of groundmass occurs in some brown coals of the Moscow region. Bands of the transparent substance are sometimes found among durains from the Kizel coalfield (Gubakhin), although these almost exclusively contain the opaque material. In transmitted light, the opaque substance is black in colour, sometimes with a brown tint along its edges; in reflected light, it is a bright white with a very slight yellowish tinge. It looks like fusain, or more accurately, like the surfaces of its cell walls. On the other hand, the transparent mass cannot be distinguished from vitrain by its colour in reflected light.

Two types of durain can be distinguished, according to the type of groundmass. The first type is a real dull coal in external appearance. The second is a little more lustrous and has a finely striated structure.

Durain is subdivided according to the predominance of one or other of essential constituents:

- (1) spore bearing (macrospores, microspores and mixed types);
- (2) cuticle bearing types;



Fig. 109. Clarain in reflected light, $\times 80$

- (3) resinous;
- (4) durain from algae (belongs to the sapropelites) or durain with algae (belongs to the mixed coals);
- (5) durain made of peculiar rounded bodies of unelucidated nature;
- (6) durain made of scraps of woody tissue in the form of xylain, xylovitrain and fusain—the fusain-xylain type;
- (7) mixed durains, consisting of two or more of the elements listed above.

The majority of Paleozoic coals prove to be spore-bearing durains. Examples are the Lower Carboniferous coals of the Moscow, Kizel, Karaganda, Pechora and Spitzbergen coalfields.

The dull *attrital* coals cannot be distinguished from durains by their appearance and physical properties; they have a complete, or almost complete absence of structureless, cementing material (groundmass), and belong to the liptobioliths, i.e., to residual coals formed from stable plant remains, mainly resin bodies.

Clarain is a complex ingredient of coal, like durain consisting of a groundmass and essential constituents, but in reverse, the groundmass predominating, while cortical and wood tissue derivatives are among the structural elements (Fig. 109). In all cases, a humic substance transparent in thin section predominates over the other constituents, i.e., over bituminous bodies and opaque material. As in the case of durain, it can occur as thick intercalations, but sometimes builds an entire coal seam. Such seams of semi-bright coal can enclose small lenses of fusain and vitrain, and intercalations of dull coal, i.e., durain (Fig. 110). Bedding is more or less discernible in it. A gradual transition to dull coal is sometimes observed. Clarain and durain predominate in the mass of coal, and there may be a durain

(dull) coal, a clarain (bright) coal, or, finally, a mixed coal (an intermixture of dull and bright).

A study of bright coal from various regions of the U.S.S.R. records several types of clarain. The most common are:

(1) Spore and cuticle clarain with a predominant humic ground-mass in which are scattered the individual constituents. Some of the semi-bright coals of the Moscow region are spore-bearing clarains.

(2) Fusain-xylain clarain, consisting of scraps of altered plant tissues in the form of xylovitrain, xylain and more rarely as structural vitrain and fusain.

Vitrain is the name given to a homogeneous ingredient of coal without any inclusions (Fig. 111). It usually occurs as lenses, sharply demarcated from all the other constituents, and is easily separated from them (as also is fusain). In most cases, the thickness of these lenses fluctuates between 0.5 and 5 cm, rarely between 10 and 20 cm. Vitrain lenses always possess a regular jointing normal to the bedding (i.e., vertical) and are therefore easily broken up. Conchoidal fracture is another of their distinguishing features.



Fig. 110. Banded coal. The black bands are vitrain, the grey, durain



Fig. 111. Vitrain with contraction cracks, $\times 50$

Emphasis is sometimes laid on the fact that vitrain bands are structureless and that they are identical with the groundmass of clarain and durain, and that they are of colloidal origin. It should be noted in this connection that contraction cracks characteristic of pure colloids can be seen in the vertical jointing cutting the vitrain bands. The latest studies, however, compel us to no longer regard a structureless ingredient as an exclusive diagnostic feature of vitrain.

Some authors define vitrain as a homogeneous substance consisting of ulmin compounds, which fill the plant cells and have partly, if not wholly, replaced the cell wall material. Similar phenomena are observed in young coals of the Baikounur, Togusken, Kok-Yangaka, Chelyabinsk and other deposits.

1. THE METAMORPHISM OF COAL

It was long supposed that any change in composition and structure of coal depended on the time which had elapsed since its formation. Detailed investigations of coals from different coalfields and of different ages have shown that this idea cannot be sustained. At the same time, the tremendous importance of temperature and pressure in determining the quality of coal has been recognized. Temperature causes the most definite effect on *contact metamorphism*. Some deposits of the Tunguska coal-bearing region have been studied more than others from this aspect.

Intrusive bodies of considerable size have been injected here into a coal-bearing series, and their effect on the coal extends to approx-

imately 2.5 km from the contact. For the first 5-10 m, the coal is converted to graphite; at a distance of approximately up to 200 m, it is altered to anthracite and hard coal with a yield of volatile components up to 18%; still further from the contact, the loss of volatiles rises to 24-25%; sometimes the coal is converted to a natural coke, suitable (but not always) for metallurgical purposes.

Although contact metamorphism is very clearly manifested, it is of limited importance, and local in character. Its effect on coal deposits is of a negative character, since it leads to partial destruction of the coal, or to a sharp alteration in its properties.

Regional metamorphism is of the greatest importance in the alteration of coal. It is associated with a lowering of the coal seams into zones of high temperature and pressure, leading to an impoverishment of the coal in volatile components. Sometimes the release of volatiles proceeds so regularly, that its gradient can be established (1-3% per 100 m of stratigraphical depth). The alteration on the release of the volatile components, for example in the Prokopyevsk district of the Kuznetsk coalfield, is given as 1.8% per 100 m of stratigraphical depth. In the largest coalfields of the U.S.S.R., where the thickness of the coal-bearing and overlying sediments reaches 10-20 km, the effects of regional metamorphism are recognized practically everywhere. Since the most intense downwarping and the greatest thicknesses of sediments belong to the central parts of the geosynclinal basin, a zonal arrangement of the different kinds of coal is seen in regional metamorphism, the coals becoming less metamorphosed towards the platform, passing from the most metamorphosed, from anthracite, for example, to the long-flamed and even to the brown coals.

2. THE EFFECT OF METAMORPHISM ON THE INGREDIENTS AND ESSENTIAL CONSTITUENTS OF COAL (AFTER YU. ZHEMCHUZHNIKOV)

The **fusain** in long-flamed and gas coals proves to be just as opaque in transmitted light, as in hard coal and anthracite, where the whole of the coal is opaque. The difference is seen only in reflected light. In coal of a lower degree of metamorphism, fusain is distinguished from the surrounding humic material by distinct high relief and a brighter yellowish-white tint; this latter forming a contrast with the greyish colour of the groundmass. In hard coals and anthracites, fusain no longer stands out by its relief, but passes directly into the unbroken field of the vitrified mass. The distinction in colour also disappears, since at this stage of the metamorphism, both clarain and vitrain acquire a yellowish tint discernible in some anthracites, even to the naked eye.

Durain is a complex aggregate consisting of groundmass and different essential constituents. The results of the metamorphism

of each of its constituents must therefore be considered, before deciding on the alteration which characterizes the aggregate as a whole.

Coals of a low-grade metamorphism have a groundmass which is translucent and brown and yellow in colour. When the metamorphism is of a higher grade, the coal changes its colour from a brown-red to dark-brown and dark-red. Hard coals do not give transparent sections, and consequently, the groundmass in them is opaque.

The humic material has a grey shade in reflected light, and the darker it is, the more transparent it is in thin section, i.e., the less metamorphosed is the coal. In brown coal, it is darker than in the long-flamed, and darker in the latter, than in the gas coals. The groundmass of this type is lighter (milk-white) in coking coal, than in gas coal. When hard coal is approaching anthracite, it begins to acquire shades of yellowish-white.

The opaque groundmass in long-flamed and gas coal is slightly translucent along the edges of the separate masses. This becomes imperceptible on metamorphism. In reflected light the opaque mass stands out with its bright white colour among the grey humic material, i.e., it is the same as fusain in colour, but is distinguished from it by an absence of relief. When the degree of metamorphism is higher, it becomes difficult to distinguish it from the transparent material, and in anthracite in ordinary light, both forms of groundmass become difficult to distinguish. The colour of both is a bright yellowish-white. The difference is made apparent only between crossed nicols.

The essential constituents. Spores. In reflected light in normal coal (little metamorphosed), macrospores and microspores are grey and dark grey and always appear darker than the groundmass surrounding them.

In transmitted light, the macrospores appear yellow and orange. The latter may be more characteristic of metamorphosed coal (for example, the gas coal of the Kizel coalfield), having more distinct outlines in brown coal, insensibly altering in the first stages of stone coal. But subsequently, the spore exines become still darker, and lose their sharpness of outline. In hard coal and anthracite, they are already as opaque as humic material and cannot be distinguished in thin sections. Besides this, some relief can be discerned even in spores, especially with oblique illumination or immersion. It also becomes gradually less distinct with increasing grade of metamorphism. There is evidently also a chemical alteration of the material of the spores, tending towards humic material.

Cuticle. Cuticle is well preserved only in the first stage of metamorphism (Π grade coal), and starting from the gas coals suffers considerable alteration, resisting the increasing effect of metamorphism far less than spores and pollen. The alteration features seen in section are similar to those described in the case of spores.

In coking coals, it is already quite difficult to identify cuticle, while in hard coal it is impossible, since here the colour of the cuticle is very near that of the groundmass.

Algae have still less resistance to metamorphism and cannot be distinguished even in some gas coals. It is especially necessary to bear this in mind when sapropelites, or what are known as pseudo-cannels, are found in coals of an intermediate grade of metamorphism.

In the highest grades of metamorphism, beginning with coking coal the presence of sapropelites cannot be demonstrated either by chemical composition or microscopic structure.

Resinous bodies in brown and long-flamed coals have a lemon-yellow or orange colour in thin section, and grey in polished section. Their transparency however gradually disappears with increasing degree of metamorphism of the coal. They become the same orange or dark-brown as the groundmass and have already lost their transparency in the coking and hard coals.

There is reason to state that some forms of resinous bodies become harder in the process and stand out in higher relief as so-called rounded bodies, for example, in Suchan coal. This suggestion, however, requires checking by further investigation.

All the constituents of durain are thus altered during metamorphism. They proceed in one general direction and can be revealed in both transmitted and reflected light. The latter is the more important in that it allows observations to be made in all stages of metamorphism right up to anthracite.

The metamorphism of coal smooths out the characteristic features of the individual elements and leads to an increasing homogeneity of the material. This refers to the optical as well as to the physicochemical features. Both types of anthracite groundmass become indistinguishable from one another, and the essential constituents become very similar to the groundmass, as if melting into it. As a result, durain appears less and less among the other ingredients of the harder coals. It must, however, be noted that etching can reveal plant tissue patterns, spore shapes and cuticle figures even in anthracite and hard coals.

When hard coals and anthracites are examined with the naked eye, we find bands which are duller as a result of a granular texture. Chemical analysis shows that this granularity is associated with a sharp rise in ash content. In slice they reveal little grains of quartz or lenticles of argillaceous material. These bands can in all probability be accepted as a durain, the groundmass and essential constituents of which have been converted into a substance indistinguishable from bright coal; but the mineral matter characteristic of durain, and not of bright coal (clarain and vitrain), does not disappear, but on the contrary, enriches these intercalations since in so doing the surrounding mass has been somewhat impoverished.

Clarain. What has been said regarding the transparent ground-mass of durain is also completely applicable to clarain, which in the highest grades of metamorphism is almost indistinguishable from vitrain in reflected light.

Vitrain, as also clarain, is determined in polished sections of long-flamed coals by its grey tint; in hard coals they acquire a brighter white and even almost a yellowish shade. Macroscopic diagnostic features of coals are given below:

	Brown coal	Stone coal
Features	Mostly brown, rarely black	Mostly black, rarely brown
Colouration with KOH solution	Quite dark	No reaction
Colouration with dilute nitric acid	From bright yellow to red-brown	Ditto

Brown coals are usually more hygroscopic and contain from 10 to 30% of moisture; in unweathered coal it is not more than 5-7%.

According to Gruner's classification, coal is divided thus: dry long-flamed giving 50-40% of volatile components (Group I); gas 40-32% (Group II); rich, or forge, 32-26% (Group III); coking 26-18% (Group IV); hard, or semi-anthracitic, 18-10% (Group V); anthracite, 10-8% (Group VI, added later).

Within this range of coal, the C content gradually increases (from 75 to 93%), and the O content decreases (from 19.5 to 3%), and the ratio $\frac{O+N}{H}$ diminishes (from 4.3 to 1%).

Donetsk coal has been graded according to Gruner's classification, but departs from it in some respects:

Grade of Donetsk coal	Yield of volatile substances, %
Д	42
Г	35-44
Ж	26-35
К	18-26
ОС	12-18
Т	<17

In metamorphic grade, anthracite is the highest member of the series among the humic coals. Its specific gravity (1.4-1.7) is higher than that of ordinary coal. It has the greatest hardness, 2.0-2.5.

The colour is greyish black. The lustre is metallic with a yellowish tinge. The fracture is conchoidal. The yield of volatiles is small. The coke does not clinker and forms a powder. X-ray investigation shows that some orientation of particles takes place in anthracites, leading to a transition from an amorphous substance to a crystalline, and the partial graphitization of the coal material, i.e., the appearance of elementary carbon.

3. MINERAL MATTER IN COAL

A certain amount of mineral grains and argillaceous material are always present in mineral coals. The minerals are usually pyrite, marcasite, quartz, and to a considerably less extent, grains of feldspar and calcite, and flakes of mica. The sizes of these mineral grains are not always the same even within the confines of one thin section, and their sorting also differs: relatively coarse and fine grains are found, often poorly or not at all rolled, but sometimes well rolled and sorted. They are usually fairly uniformly distributed in the groundmass, do not form distinct layers and are stratified only in certain coals.

The argillaceous material found in coal and carbonaceous rocks is nearly always kaolinitic. Under the microscope, in transmitted light, the clay, if not contaminated by the organic component, is most often yellow and only sometimes of a greyish shade. The yellow argillaceous material of colloid type is absolutely uniform, with and without the polarizer. The argillaceous material of greyish tint is also very uniform without the polarizer, and only occasionally at high magnifications it is possible to distinguish mineral particles in it, but they are more distinct in polarized light. In such cases, the grey clay is distinguished from the yellow by the very fine-grained structure, and the very small grains comprising it have a different interference tint.

One of the essential characteristics of coals is their ash content (the solid residue formed when coal is burnt). The colour and type of the ash of the constituents of coal are not the same.

Fusain—dark grey-brown, compact residue
Durain—pale grey, almost white powder
Clarain—reddish-brown fluffy residue
Vitrain—pale yellow light residue

The ash content differs in brown and stone coal, from 3-4 to 40-45%.

The main constituents of ash are silica, alumina (20-40%) and ferric oxide; sometimes the latter attains 60-75%. The lime content sometimes rises to 10-19%. The remaining constituents (silicates, sulphates, carbonic acids, sulphites and phosphates) are present in amounts ranging from fractions of one per cent to several per cent.

Besides this, certain rare elements are present, but can usually only be detected by chemical, spectrographic or X-ray analysis. Their greatest concentration is in the ash of low-ash coals. According to Goldschmidt, the presence of the following elements has been established in coal ash:

Be, Ca, Sr, Ba
B, Sc, Y, La, Al
V, Co, Ni, Mo, Pd, Rh, Pt, Mn, Fe
Cu, Zn, Ga, Ge, As, Se, Ag, Cd, Sn, In, Au, Pb, Bi

The following contents are known in some coals:

Ag = 1.3-10 g/ton; Au = 0.5 g/ton; Pd = 0.05-0.2 g/ton;
Pt = 0.1 g/ton.

The amount, character and distribution of this mineral matter is one of the most important causes of the change in the physical properties of coal (colour, lustre, specific gravity). Coals with more ash are heavier than those with less ash. The ash content increases the dullness of the coal, a strongly contaminated clarain, for example, losing its inherent brightness. An increase in the ash content of bright coals entails a decrease of jointing, since a uniform distribution of mineral particles renders the coal less brittle.

A finely dispersed arrangement of the mineral particles turns out to have a greater effect on the groundmass.

The properties of ash (Table 12) are not equivalent as regards solubility, and are related in different ways with the material of

Table 12

Properties of Ash

Form of ash	Properties	Removal on concentration
Constitutional-vegetable (alkaline)	Mostly dissolves in water	Not removed
Internal-infiltrational (carbonates and sulphides)	Dissolves in HCl	Partially removed (as coarser inclusions—pyrite nodules, gypsum films, clay lenticles, etc.)
Internal-imported (clay and sand)	Insoluble in HCl	
Extraneous—from soil, roof and intercalations	Ditto	Completely removed as mineral admixtures from intercalations, soil and roof

which the coal is made up, and, as a result, it may be correspondingly easier or more difficult to remove them by concentration methods.

The clay minerals, in the first instance, kaolinite, and pyrite and calcite, are the main mineral substances in intrinsic ash; they have been brought into the coal seam by infiltration in the course of accumulation of the peat. The clay had been brought to the swamp by running water, and the sulphides and calcite were formed in the coal joints, in coal balls and cavities and are of secondary origin.

The origin of finely disseminated pyrite can probably be attributed to the activity of sulphur-forming bacteria. The larger masses of it (sometimes even a band up to 30 cm) are due to the deposition of iron sulphate when the hydrogen sulphide obtained from the decomposition of plant material interacts with iron compounds dissolved in the swamp water. It is possible that some pyrite may have been formed by reduction of ferrous sulphate by organic substances.

Coal lying near the surface can contain clay, hydrated ferric oxide, ferrous carbonates or sulphates brought by percolating surface waters and deposited in jointing and cleavage cracks.

Free sulphur is not found in coal, but is an ingredient of organic compounds, and of pyrite or marcasite, and is also contained in weathered coals as calcium sulphate. This sulphur is termed organic, pyritic or sulphatic. It is known to have a considerable effect on the quality of the coal. It must be noted that organic sulphur is distributed throughout the entire substance of the coal, and cannot be separated by concentration. The presence of sulphate sulphur is almost a definite indication that the coal has been weathered.

As well as distinctly visible mineral grains, and bands and lenses of argillaceous material relatively free of organic additions, clay is found as a dispersion disseminated in coal.

The relation between the various mineral impurities and the genetic types of coal will now be examined.

Bright clarain coals are nearly always characterized by a considerable content of siderite. Of the other mineral additives, quartz, calcite and pyrite can be observed. The presence of this large amount of siderite in this coal, as well as the small amount or absence of quartz, is evidence of its formation in quiet, stagnant basins, in an anaerobic medium with a large CO_2 content. Siderite is, as a rule, associated with the groundmass of the coal and very rarely fills in the residues of cell cavities in plant tissues. It is mostly distributed uniformly as individual grains; more rarely, it forms a lit-par-lit arrangement of lenticular and banded aggregates.

Inclusions of calcite and an argillaceous substance are also characteristic of bright clarain coals, for these coals, being brittle, contain many cracks occupied by epigenetic minerals.

Semi-dull clarain-durain and dull durain coals are characterized by the presence of a large amount of clastic material, mainly quartz,

more rarely calcite and argillaceous material, while flakes and scales of mica are present in small amounts. Minerals of chemical origin (calcite and chalcedony) are seen in considerably lesser quantities. They fill in the residues of cell cavities and appear as inclusions in the groundmass. Siderite and pyrite are absent. The large amount of clastic material suggests that the coal was formed in running water. The absence of siderite and pyrite, as well as the presence of chalcedony, indicates the predominance of aerobic processes during the conversion of the plant material. Epigenetic minerals are not characteristic of semi-dull coals.

Semi-bright durain-clarain coals are characterized by an almost equal content of clastic (quartz, calcite and argillaceous material) and chemical (calcite, siderite, pyrite and chalcedony) minerals.

CHAPTER XIV

Metamorphic rocks

Metamorphic rocks are formed by the alteration of igneous or sedimentary rocks as a result of changes in external conditions after their initial development. These changes proceed without melting or solution of the rocks.

Traces of the original structure and composition of a metamorphic rock can be found by making a careful study of its peculiarities of structure and composition, and this provides a way of determining whether it is the product of the alteration of an igneous or sedimentary rock. The Finnish geologist Soderholm has compared these rocks with what are known as palimpsests, i.e., parchments from which chroniclers had effaced former writings in order to use them for new ones, but with the experienced eye of the investigator finding traces of the more ancient writing under the new; metamorphic rocks can also be regarded as palimpsests, and their structures be called palimpsest structures.

All sedimentary rocks are formed under conditions of relatively low temperature and pressure. In the course of mountain building, every rock can sink into the deep-seated zones of the crust, thereby naturally falling into a completely different set of conditions, as a result of which it is altered into what is now a metamorphic rock.

A simple calculation shows that the temperatures prevailing at depths between 60 and 100 km should melt any silicate rock. At the same time, the pressure also increases sharply with increasing depth of crust, being raised by one atmosphere for a depth of 2.5-3 m.

Zones of rock fracture and crushing can be assumed to arise in the upper regions of the crust, where the pressure of superincumbent beds is not yet very high. The pressure further down is so great that the directed tectonic forces are relatively feeble. At the lowest horizons, the pressure exceeds the elastic limit of a rock, and at the same time, the rock experiences the emollient effect of increased temperature. Any directed pressure is already impossible here, because of the plasticity of the rock, and only a uniform hydrostatic pressure exists.

Table 13

Distribution of Temperature and Pressure in Metamorphic Zones

Metamorphic zone	Pressure		Temperature
	directed	uniform	
Upper zone (epizone)	Can sometimes be very high	Low	Low
Intermediate zone (mesozone)	Very high	Medium	Medium
Lower zone (katazone)	Low	Very high	Very high

The alteration of rocks depends on the conditions in the different zones in which the rocks occur (Table 13).

In the epizone, the alteration is weak, the structure of the original rocks being preserved as a rule, and schistosity or foliation making an appearance only in the lowest parts of the zone where the pressure is already high enough. Since mechanical shattering predominates in the upper metamorphic zone, breccia-like products are created from granular igneous rocks.

The character of the rocks arising in the mesozone is determined by directed pressure, which orients most minerals in a definite manner, giving all rocks a foliated appearance. Schists are the typical rocks. There is no directed pressure in the katazone, and consequently no oriented mineral grains; rocks free of foliation or schistosity again appear; their very characteristic mineral composition serves to distinguish them from rocks of the uppermost zone.

In the course of mountain building, large areas of the crust can sink down to a considerable depth, and, hence, the rocks of which they are composed can suffer alteration. In this case, it is difficult to observe differences in metamorphism, even if the same horizon is followed for very great distances. This kind of metamorphism, which developed at great depths below the surface, is not confined to the contact with magmatic rocks, and leads, as a rule, to the creation of large masses of schists; it is termed *regional* or general (sometimes deep-seated) *metamorphism*.

There is another type of metamorphism which is widespread. During the injection of an intrusion, a magma heats up the enclosing rocks and alters (metamorphoses) them. There are, however, essential differences between this and the regional metamorphism of similar rocks; firstly, the part played by pressure is much less, and secondly, the intrusion is cooling, and consequently affects the surrounding rocks for a relatively short time, and therefore the zone

of altered rocks, forming a border around the intrusive, will be relatively small. This form of metamorphism bears the name of *contact metamorphism*.

The phenomena of contact metamorphism are naturally clearly distinguishable from regional metamorphism when they are developed in the epizone; but if the intrusions are injected into deeper horizons, contact metamorphism will merge into the regional and the differences between them will be disguised.

In the lower metamorphic zones, the magmatic material frequently impregnates the enclosing rocks or infiltrates along the schistosity; the schists formed in this case are known as *migmatites*. The original material is partially preserved in their composition, but the newly formed or introduced minerals are the most significant. When fine layers of injected granitic material alternate with fine bands of the host (usually metamorphosed) rocks, the migmatite obtains the name of *lenticular gneiss*.

The process involving the complete melting of rocks at depth with the formation of new magmas capable of injection and passage through the cycle of magmatic conversions has been named *palingenesis*; some of the Archean granites and granodiorites in Fennoscandia are considered to be palingenetic. If this process takes place under the influence of hot gases ascending from below, it is called *anatexis* (ultrametamorphism).

Besides the main types of metamorphism—contact and regional—several others can be distinguished, but they are less important.

Autometamorphism is the alteration of a magmatic rock by volatiles and solutions given off in the cooling process. Its scale is not always clear; alteration processes which can be referred to this type, include the serpentinization of ultrabasic rocks, the chloritization of diabases, the albitization of spilite-keratophyres, the felsitic regeneration of the glass of acid effusives.

Dynamometamorphism is shown by a change in structure of a rock often due to the crushing of the constituent minerals under pressure (mostly from folding) without any participation of high temperatures or magma. No new minerals are formed. In its pure form, this type of metamorphism is observed only in the upper parts of the epizone, where mountain-building movements can create local stresses of great intensity, unaccompanied by a rise in temperature. The greatest changes take place along fracture lines where special rocks can be formed, consisting of finely crushed and cemented fragments (tectonites and mylonites).

Metamorphic rocks were for long an enigma to investigators, and it is only recently, due to the development of physical chemistry, that any approach could be made towards a scientific understanding of the basic processes of metamorphism.

The hypothesis of regional metamorphism was first formulated by the Russian scientist Lukashevich in his work *The Inorganic*

Life of the Earth. A similar theory was developed parallel to this by Van Hise, Becke and Grubenmann. The latter separated the three metamorphic zones: the epizone, the mesozone and the katazone. According to him, each region is characterized by its own thermodynamic conditions, and consequently, also by its own metamorphic processes, which cause the development of special typomorphic minerals.

But this hypothesis ignores such phenomena as magmatism and folding with which magmatic regions are invariably associated.

By the depth factor, Grubenmann understood a proportional increase of temperature and pressure with depth; but neither temperature nor pressure are identical with depth or proportional to it. Metamorphic formations are usually associated with mobile fold zones, and intruded magmatic masses. The temperature and pressure in these zones are independent variables and are not directly connected with the absolute depth to which the sediments have sunk.

D. Korzhinskii distinguishes in the main two types of metamorphism: (1) common, or normal metamorphism, during which the chemical composition of the rocks, with the exception of water and carbonic acid, remains constant; (2) metamorphism of rocks, involving a more fundamental change in chemical composition, called metasomatic metamorphism or metasomatism.

According to Korzhinskii and a number of other investigators, the temperature of the metamorphic process depends on the temperature and the amount of heat which is being given off by the magmatic bodies, and also on the conditions of heat transfer, which in virtue of tectonic disturbances differ in various parts of the rocks, even if they are equally remote from the massif.

When a magma intrudes sedimentary beds, it loses its heat during its migration through them, and if its mass is relatively small, it congeals in the form of porphyritic rocks with a glassy groundmass independently of the depth at which this takes place, and conversely, when the magma masses are large, coarse-grained magmatic rocks can be formed at relatively shallow depths.

The concept of depth is not synonymous with either the concept of temperature of any one region, or with the concept of the limits of partial pressure. It cannot, all the more, be synonymous with the concept of the grade of metamorphic processing of rocks, depending as this does to a great extent on the mobility of the material and the degree to which it is impregnated by solutions. These phenomena depend on the intensity of the tectonic and magmatic processes.

The magmatic processes which are the main cause of metamorphism follow different courses in regions where the mobility and the thickness of sediments accumulated are different, and also at different stages of their development. It is therefore natural that metamorphic processes proceed differently in regions of different structure.

Regions of extensive development of the most deeply metamorphosed series are usually associated with geosynclines. Great masses of magma are injected here, and this means that tremendous masses of rock are heated up.

In the transitional zones between geosynclines and their platforms, the accumulation of sediments is less, and the magmatic masses are smaller in volume. The metamorphism here develops around the massifs in the form of contact aureoles with transitions from the hornfels grade to metamorphic schists and little altered normal sediments.

In platform regions, magmatic injections appear as dykes and flows of sheets and sills; the metamorphic phenomena are confined to local pyrocontact actions on normal sedimentary rocks.

The supplies of thermal energies and solvents brought by the huge magma masses are sufficient to heat up large volumes of rock material, and the reaction aureoles of these masses extend over great distances and are therefore regional.

Magmatic phenomena are the necessity in the metamorphism of silicate rocks for they introduce both energy and solutions. It must be concluded from this that regional metamorphism is in many cases merely concealed contact metamorphism. There are cases where a powerful metamorphism is seen in the vicinity of comparatively small magmatic injections. Intensive metamorphism is sometimes discovered at quite a distance from a massif without any obvious link with it. This dissimilarity in the intensity of metamorphism of magmatic masses is explained by the fact that it is controlled not only by the size of the magmatic body, but also by its temperature and degree of mechanical mobility, as well as by the permeability of the enclosing rock masses towards the solutions circulating in the zone of metamorphism.

The hypothesis of the deep-seated belts of regional metamorphism regards the process of rock alteration as an event which takes place in situ under the action of varying loads; meanwhile, every characteristic feature of metamorphic rocks—cleavage, schistosity, the linear orientation of the minerals, porphyroblasts and rotation structures around them, banding and the folded structures due to traction—is evidence that metamorphic rocks did not crystallize in motionless beds, but in a series of rocks in mechanical movement, being displaced during a period of folding. And, finally, the sum total of metamorphic phenomena, as of any phenomena taking place in the crust, cannot be regarded as a reversible process. Even in the case of repeated metamorphism of one series, the process will proceed in what is basically a new medium and in new conditions.

Metamorphism is sometimes characterized by repetitivity, and when making a study of it, metamorphic phases and the phenomena characteristic of each phase must be regarded separately.

A detailed investigation will always succeed in establishing that each metamorphic phase corresponds to a definite phase of tectonic activity.

1. ROCKS FORMED BY CONTACT METAMORPHISM

When a granite magma acts on slates situated at some distance from the massif, segregations recognizable to the unaided eye make their appearance consisting of carbonaceous material or graphite, micas, scales of chloritic material, ores or other newly formed minerals; the main mass of the rock may remain unchanged, spotted or knotted slates then being formed.

In the zone nearer the intrusion, the original argillaceous material of the same slates has been recrystallized. Quartz and mica predominate in it. Individual crystals of andalusite, cordierite and feldspars occur along with the dark segregations. The rocks thus formed are andalusite-mica contact hornfelses or similar rocks.

In the immediate vicinity of the contact, the spotty segregations disappear, and also the original bedding or foliation of the rock, and compact dark hornfelses are formed: these consist predominantly of feldspars, andalusite, cordierite, magnetite, biotite and quartz, with a well developed though fine granoblastic texture.

When basic rocks, for example gabbro or diabases, act on the slates at their contact, the spotted slates formed are spilositites with green or greyish-green (rusty on weathering) segregations of chlorite, or adinole, which are compact, fine-grained, sometimes schistose rocks, consisting mainly of albite and quartz.

Silicates rich in calcium and iron (diopside, epidote, garnets, wollastonite, etc.) are formed from solutions in zones of contact alteration when an acid magma acts on carbonate rocks: these are known as skarns. Skarns are, in most cases, characteristically metasomatic rocks and are formed along the path of metamorphosing solutions, which are introducing silica into the carbonate rocks, or lime into the zone of silicate rocks. The formation of skarns at the contacts of silicate and carbonate rocks occurring in a region subject to the action of solutions is very typical; an exchange of material takes place here under the influence of the solutions, giving what Korzhinskii called bimetasomatic skarns. Skarns are often associated with certain ore deposits (magnetic iron ore, copper ores, polymetallic ores, tin and tungsten).

2. ROCKS FORMED BY REGIONAL METAMORPHISM

Directed (unilateral) pressure lowers the melting point of minerals, and increases their solubility, thus representing a powerful factor in recrystallization; solution therefore begins in regions where there

is the greatest pressure on the crystal with parallel deposition of dissolved material in regions of least pressure. This phenomenon provides a satisfactory explanation of the origin of the parallel schistosity and foliation structures in metamorphic rocks. Crystals will always grow in the direction perpendicular to the greatest pressure, and develop elongated forms oriented parallel to the direction of least pressure. Directed pressure thus appears as the main cause of the parallel structures and textures so characteristic of metamorphic rocks.

Unilateral pressure, which is seen to be a predominantly comminuting agent in the upper horizons of the crust, becomes a creative process with increasing depth, promoting the formation of certain minerals which are especially stable under rock pressure (biotite, muscovite, chlorite and amphiboles, kyanite, staurolite, chloritoids, talc). The practical importance of this fact is particularly great because there exist a number of minerals which are not developed in crystalline schists, but are well known as products of contact metamorphism. These are basic plagioclase, olivine, andalusite and others.

The action of oriented pressure on rock and mineral fragments appears mainly in the form of curvature, undulose extinction and the development of a mosaic structure in quartz, the appearance of various secondary twinning in crystals of feldspars and calcite, and optical anomalies in isotropic minerals.

If the crystalline schists resulting from pressure (regional metamorphism) prove to have been displaced from one metamorphic zone to another, their mineral composition and structural features will in the course of time adapt themselves to their new circumstances. This process is called *regressive metamorphism*. Thus, in the Alps, some of the phyllites and fine-grained mica schists have been formed from deep-zone gneisses.

Phyllites are typical metamorphic rocks of the epizone, being argillaceous mica schists, or micaceous slates. They are finely bedded, finely foliated black or dark grey, sometimes greenish (chlorite, talc) rocks, sometimes with shades of violet, and consist of fine-grained scale-like fragments, mixed with newly formed clay particles, quartz and mica, and as a consequence of the large content of fine flakes of mica (sericite) they have a specific silvery hue. Phyllites gradually pass into ordinary slates, which are the first stage in the metamorphism of shales.

Crystalline schists are typical of the mesozone. They are holocrystalline schistose silicate rocks, which can be formed from igneous (orthoschists) and sedimentary (para- and metaschists) rocks. Many of the minerals forming crystalline schists are typical components of igneous rocks. But along with these are found others which are not formed when an igneous rock crystallizes; such, for example, are staurolite, kyanite, sillimanite and other minerals and specific



Fig. 112. Secondary growth of quartz in quartzite, $\times 140$

mineral associations which can be created only under metamorphic conditions.

Although the crystalline schists also possess a holocrystalline texture, they never show such a definite sequence of crystallization as do the igneous rocks. The term *crystalloblastic* has been adopted for the texture of crystalline schists, indicating that they were formed by the growth of crystals in a solid medium by means of recrystallization, and not by transferring material from the liquid to the crystalline state. Crystals which have preserved their idiomorphic outlines under these conditions are termed *idioblasts*, phenocrysts are called *porphyroblasts*, and crystals with no definite shape, *xenoblasts*. Not infrequently, there are prominent bands or segregations composed of some mineral or mineral association. In those cases when the original minerals and structures are still partially preserved and are seen traced out against the general background of new crystalline formations, we speak of *relict structures*.

Mica schists consist mainly of quartz, scales of mica and chlorite. They are distinguished from gneisses by an almost complete absence of feldspar. They have a more or less well developed bedded structure, especially well presented in varieties rich in mica. Muscovite, biotite and paragonite schists are distinguished by the type of mica, or feldspathic, garnet, staurolite and other schists by their secondary constituents. A relative abundance of talc and chlorite allow some schists to be known as talc schists or chlorite schists. When the content of quartz increases, mica schists pass into micaceous quartzites and quartzite schists.



Fig. 113. Gneiss. The Northern Caucasus (Digoriya)

Quartzites are pure or almost pure quartz sandstones which have been converted by metamorphism into granoblastic quartz rocks, usually with a small amount of mica, tourmaline, graphite or iron minerals (Fig. 112).

They sometimes possess a brecciated or schistose structure. The colour of quartzites is usually grey of various shades down to black (tourmaline, magnetite), or even red (hamatite). The same name is given to a fine-grained quartz sandstone with a quartz cement. The Shoksha quartzites are especially well known in the U.S.S.R. (from the western shore of Lake Onega), as also are the iron quartzites of Krivoy Rog, which are rich in magnetite, martite and other iron minerals.

Amphibolites are granular or schistose dark-green, essentially amphibole rocks. Plagioclases occur in conjunction with the hornblende mineral in feldspathic amphibolites. If the garnet begins to predominate in garnet-bearing amphibolites, the result is a garnet amphibolite.

Gneisses (Fig. 113) are the typical rock of the katazone. They have a foliated structure, and consist of feldspars, quartz and a coloured mineral. They thus correspond to granites in composition, although they are often metamorphosed sediments in origin, as is shown by their chemical composition (a large content of alumina; predominance of magnesium over lime, of potassium oxide over sodium oxide), remains of the original bedding, and the outlines of former pebbles

which are met from time to time. But gneisses are known which are the result of the metamorphism of granite, adamellite or quartz diorite. The first are called paragneisses, the second orthogneisses. The gneissose appearance can be obtained in granite rocks during crystallization and consolidation at great depths, when strong hydrostatic and directed pressure are exerted at the same time.

Gneisses are distinguished by the predominant mineral composition: biotite, hornblende, two-mica, and so on. Mica schists are closely related to the gneisses by gradual transitions, and are distinguished from them by a greater amount of mica and quartz, and also by the more distinct schistosity.

Granulites are light, compact, fine-grained, schistose gneiss-like rocks consisting of feldspar, quartz and garnet, with biotite, hornblende and augite. They sometimes form massifs tens of square kilometres in area, and surrounded by an envelope of phyllite and mica schist.

Marble. In all metamorphic zones, limestones suffer recrystallization and are converted into crystalline limestone or marble.

If the limestones contain some siliceous and argillaceous material, or if the oxides of silicon, aluminium, iron and some easily mobile compounds of other elements act upon them in the form of magmatic emanations, which make frequent appearances particularly at the contacts with magmatic rocks, the limestones react with them with the formation of specific minerals (diopside, phlogopite, zoisite, tremolite). Under the same conditions, dolomites experience dedolomitization, during which the dolomite is reduced to calcite, while its magnesia enters into new minerals, such as, for example forsterite (Fig. 114). Another feature of marbles is their very weak development of schistosity. Due to the plasticity of calcite, even a relatively small directed pressure passes into a hydrostatic

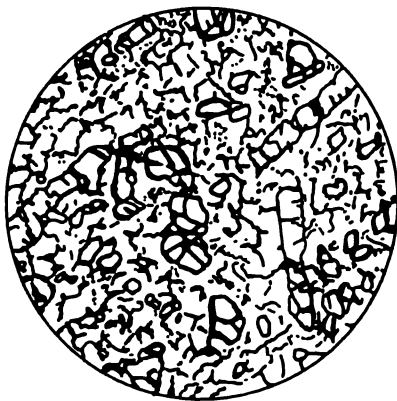


Fig. 114. Forsterite marble, $\times 20$

pressure and cannot be expressed in the orientation of this mineral. Marbles are therefore usually massive, and have an equigranular texture; it is only occasionally that they are roughly bedded and show a tendency to schistosity.

Marble, which contains serpentine diversely coloured, gives beautifully coloured serpentine marble or ophicalcite.

Technical importance of the main mass of metamorphic rocks is less than that of igneous and sedimentary rocks. The schistosity and possibility of fine splintering associated with it prevent the utilization of most schists as building material. At the same time, marbles, quartzites and some gneisses, in which the schistosity is not so strongly developed, are used for construction work and even as facing stone.

CHAPTER XV

The origin of magma and magmatic rocks

The size and scope of this textbook will not allow us to dwell on this great and complex question in any detail. Only the most important facts, and the most widely-held views concerning the origin of magma and the phenomenon of granitization will be dealt with in this chapter.

It was pointed out in the survey of the magmatic rocks, that there is a point of view which affirms the existence of only one parental magma, the basaltic. At the beginning of this century, the supporters of this view explained the formation of granite as the result of crystallization differentiation of a basaltic magma. Subsequent work showed that this idea could not be sustained, if only as a result of the facts that: (a) granites are tremendously widespread, but their transition to basic rocks is extremely rarely seen; (b) there has been no success in creating experimental conditions under which the last phase of the consolidation of a melt would have a granitic composition.

New ideas as to the formation of granite were therefore required. Scandinavian geologists suggested that the solution of the question be sought in granitization. According to their ideas, granitic (or rather, gneissose) rocks are the result of the action of emanations on crystalline schists in the deepest parts of the crust. These emanations are alkaline and rich in silicic acid and alumina, and rise up from magma, most probably basaltic, occurring below. Injection gneisses and migmatites can be created in this way, and when they are remelted, a secondary "granite magma" is formed, now able to give real granitic intrusive rocks.

In spite of its beauty, many objections have been laid against this "granitization hypothesis". In particular, no satisfactory reasons have been given to explain why granites predominate among intrusive rocks, why acid emanations are given off by a basaltic magma, or why the migmatite melts a second time.

Discussion of the granitization hypothesis has made it clear that there is a connection between metamorphism and the formation

of magmatic rocks, and that the solution of sedimentary and metamorphosed material in a magmatic melt plays a very important part in the formation of certain types of rocks.

Investigators of granites can at present be divided into two groups: the defenders of the idea of the magmatic origin of granites, and the transformationists, supporters of the conception that the purely metasomatic granitization process plays a leading role in the formation of large granite intrusions.

The substance of the transformationists' ideas can be reduced to the statement that granites and granitoid rocks are generally recognized to be not strictly granitic, but essentially metamorphic products. They arise from a metasomatic treatment of a rock of any origin and almost any composition (and not only from crystalline schists). In the general case, this metasomatic granitization is accompanied by the introduction of K, Na, Si and in some cases the removal of Mg, Fe, Ca.

The transfer of material during the granitization process is accounted for in various ways: by the infiltration of liquid or gaseous granitizing solutions; by the diffusion of ions in a solid medium; by the diffusion of ions in a liquid medium (immobile)—in intergranular liquid films.

There are also substantial differences of opinion concerning the question of the independent existence of an ultrabasic magma. Many investigators consider that all ultrabasic rocks are the products of crystallization differentiation of a basaltic magma. Others consider that ultrabasic rocks are formed from a special ultrabasic magma, which arises on the melting of the peridotite substratum. A third group recognizes both genetical types.

S. Chetverikov studied the structure and composition of an artificial rock formed from the fusion of marls when coal was burnt underground, and came to the conclusion that basic magmas can be the products of selective melting of any basic material, and that it is not necessary to believe in the presence of a basaltic shell in the lithosphere. V. Lodochnikov had come to the same conclusion much earlier, using the results of investigations of natural rocks.

It can now be regarded as established that many, if not the majority of groups and even the forms of magmatic rocks can be produced by independently existing granitic, basaltic or ultrabasic magmas, and can also be the products of granitization, i.e., essentially metamorphic formations. The diversity of magmatic rocks, arising from each type of original magma, is due, in the first instance, to the phenomena of assimilation and differentiation. Enough data have now been accumulated to speak of the similarity of features of magmatic rocks of various origins as absolutely usual and normal phenomena, and to assume that the totality of rocks usually accepted as magmatic are a complex of formations which are rather varied genetically.

Important considerations concerning the causes of the diversity of igneous rocks have been expressed by D. Belyankin. Starting with the results of studies on slags, he regards the possibility of differentiation directly in a liquid magmatic melt as doubtful, and shows that it is not the differentiation of liquid slags, but, on the contrary, the gradual levelling up of the original chemical heterogeneities that is a very usual phenomenon in them. However, it must be taken into consideration, that the conditions for the production of slags, about which Belyankin was writing, are very different from the conditions in which magmatic rocks were formed in nature.

Contact phenomena arising at the boundary of two or more media with different chemistry and different states of aggregation deserve special attention.

There are great possibilities here of obtaining a whole range of quite diverse rocks, including rocks of unusual composition, and of a composition which differs sharply from that of the active magma.

Great importance must be paid to mobile and particularly to the easily volatile substances: thus, water, for example, dissolves magmatic material selectively and bears it away to new levels, where it is fixed into the composition of new types of rocks.

Tectonics are of great importance in the creation of magmatic rocks. A variety of geological formations are brought into contiguity during any geological process; the external conditions controlling the physico-chemical processes taking place in the crust, i.e., the pressure and temperature, alter. This all evokes the appearance of a diversity of magmatic rocks or opens the way for the advancement of the magma and its derivatives.

CHAPTER XVI

Experimental and technical petrography

The laws of crystallization of artificial silicate melts of accurately known composition have been studied for some time in a number of physical chemistry laboratories and laboratories of experimental petrography, and these investigations are exceptionally important for contemporary petrography, particularly for an understanding of crystallization phenomena in natural media.

Such investigations have in view first of all the detection of the limits of pressure and temperature within which each melt is in equilibrium with its crystals, and then the elucidation of the nature of the primary crystallizing solid phases and their stability fields.

Systems are divided into one-component, two-component, three-component systems, etc., according to their number of components.

Let us consider two cases of equilibrium in a two-component system. Equilibrium here depends, besides pressure and temperature, on the concentration x of one of the components (it will be $100-x$ for the second component).

(1) *The eutectic melt.* Let us construct the solidification diagram for a liquid mixture (Fig. 115), with concentration along the axis of abscissae, and temperature (t) along the axis of ordinates, with p

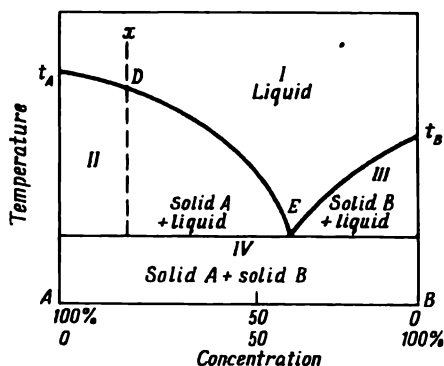


Fig. 115. Solidification diagram of a liquid mixture

constant. The points *A* and *B* correspond to the pure components *A* and *B*, t_A and t_B are the melting points of the pure components *A* and *B*. The lines $t_A E$ and $t_B E$ give, respectively, the melting points of the substance *A* with increasing amounts of *B*, and of *B* with increasing amounts of *A*. Both curves intersect at the point *E*, which is therefore the lowest temperature, at which a mixture of *A* and *B* can melt. This point is called the eutectic, and the composition corresponding to this point, the eutectic composition.

In field *I*, there is only one phase (an absolutely homogeneous liquid), and in it, concentration and temperature can be changed arbitrarily, and the liquid (phase) composition will not alter.

If a given composition (*x*, for example) is cooled, at first, there will be no changes in the phase equilibrium. Crystallization begins only after the point *D* on the melting point curve $t_A E$ is reached. The pure substance *A* first separates, and as a consequence, the residue is enriched in the substance *B*; on further crystallization, the composition of the melt should therefore be displaced along the line *DE*. When the temperature is lowered slowly, the separation of *A* continues without interruption until the residual liquid reaches the eutectic state at point *E*. At this point, *B* will also be in equilibrium with the melt, along with *A*, i.e., the crystallization of *B* will begin. We shall then have three phases: two solid, and one liquid. Temperature and composition are bound to remain unaltered at this point, until everything has solidified. When the composition of the melt is that of the eutectic, the liquid phase is converted directly to the two solids, *A* and *B*, which crystallize together and give a eutectic structure of mutual intergrowth. For compositions to the right of *E*, *B* separates first, then the eutectic.

The crystallization of systems of three or more components proceeds in a similar way, with the difference that ternary and quaternary, as well as binary, eutectics occur.

The sequence of crystallization of minerals in rocks, already referred to, and established on the basis of the idiomorphism of the minerals, can be satisfactorily explained by reference to eutectic diagrams.

Thus, in the fusion diagram of the system diopside-albite, albite possesses the lower melting temperature, and so the eutectic point is shifted towards albite, and there is a much greater probability that diopside will be the first to crystallize, which does, in fact, correspond to what is observed in nature.

Along with the eutectic, the widespread phenomenon of incongruent melting, i.e., melting with decomposition, plays an important part in the crystallization of silicate melts and magmas. Enstatite can be taken as an example: when heated to 1557°C, it dissociates into a silica melt and forsterite crystals:



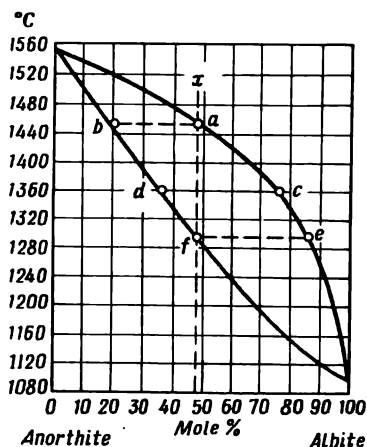


Fig. 116. Fusion diagram of the plagioclases

The same phenomenon affects the relationship of the olivine and pyroxene crystals in gabbros and basalts. At temperatures above the incongruent melting of pyroxene (enstatite), olivine (forsterite) crystallizes; when the melt is cooled below this temperature, olivine becomes unstable and begins to melt, while pyroxene crystallizes in its place.

(2) *Solid solutions.* When isomorphous mixtures are formed, the composition of the solid can change continuously throughout crystallization.

The plagioclases are a well known example, among rock-forming minerals, of a system of two completely miscible substances. The melting phenomena of the plagioclases are given in the diagram of Fig. 116.

This diagram shows, for example, that a mixed crystal consisting of 50% of anorthite and 50% of albite, begins to melt at 1287°C and is completely melted only at 1450°C. On cooling from the higher temperature *x*, the same phenomena are obtained, but in reverse order: the crystallization begins at 1450°C and is complete at 1287°C. The composition of the mixed crystals in equilibrium with the melt at a given temperature can be easily ascertained from the same diagram; thus, the mixed crystal of composition *b* is in equilibrium with a melt of composition *a*, and the mixed crystal *d* with melt *c*. It follows from the shape of the curves that a crystal is always richer in lime than the melt with which it is in equilibrium.

During crystallization of a plagioclase melt, two cases can be distinguished:

(1) Crystals are removed as they are formed; the melt becomes richer in soda and finally reaches the composition of pure albite.

At the end of solidification a mixture of mixed crystals ranging in composition from *b* to pure albite is obtained.

(2) The crystals remain in the melt, which reacts with them and is changed in composition and forces the crystals to assume a composition which is in equilibrium with the melt at each temperature (in the case of very slow cooling, when there is sufficient time to establish equilibrium). The change in the composition of the melt from *a* to *e* proceeds parallel to the continuous change in the composition of the precipitated crystalline phases from *b* to *f*.

The residual melt at the moment of solidification has the composition *e*, and in equilibrium with it is the crystalline phase *f*, which has the same composition as the original melt. The final result in this case is a homogeneous crystalline phase with the composition of the original melt.

If time is insufficient for the completion of the reaction between crystals and liquid, i.e., the cooling is rapid, and the rate of diffusion small, true equilibrium will not be attained and the result will be plagioclases with a zonal structure, the inner layers being richer in the anorthite molecule than the external.

Technical petrography is a special branch of general petrography, and concerns itself with the study of the artificially prepared rocks of technology.

The investigation of the petrography of artificial stones by applying experimental and detailed physico-chemical methods was begun in the first decade of the XXth century and is still proceeding on a very wide scale.

Technical rocks resemble natural rocks to a certain extent.

Igneous rocks —	metallurgical slags, high-early-strength cement, glass, fused corundum, spinels, etc.
Metamorphic rocks —	dinas, grog, cemented clinker-porcelain (sic)
Sedimentary rocks —	concrete, cements of various kinds, silica brick, and so on
Contact formations —	zones of waste dinas, protective scale of chamotte (grog)
Pneumatolytic bodies —	new growths on the walls and brick regenerating chambers of metallurgical and glass furnaces
Inclusions in magmatic rocks —	stones in glass
Special magmatic structures —	glass, spherulites, fluctuation structures in glasses, banded dinas

These, however, are all rather specialized: their importance is that they afford investigators an opportunity of finding a way to improve the production of artificial stones, to raise their quality and widen their variety.

Technical rocks, like natural rocks, are characterized predominantly by a silicate chemical composition, as a result of which they are built up of more or less the usual rock-forming minerals, minerals of the quartz group, corundum, magnetite, forsterite, fayalite, melilite, feldspars and many others often being present. On the other hand, the relationship of the chemical elements in technical rocks is often different from that in natural rocks, and this is reflected in their mineral composition. In igneous rocks, for example, alumina, lime and alkalies occur together constantly and with silica in characteristic paragenesis. Such a paragenesis of these compounds does not, as a rule, occur in technical products, but only as a rare exception. In particular, only lime and soda combine with silica in industrial glasses, and there is an almost complete absence of alumina, whereas in the fireclay ware of glass foundries we have just the opposite relationship: a large amount of alumina and only an insignificant percentage of lime and alkalies.

Some of the minerals named above are characteristic of natural and technical rocks alike. There is also a parallel series of minerals which are completely absent in natural rocks or else a rarity. It is sufficient to name as examples mullite, $3\text{Al}_2\text{O}_3 \cdot 2\text{SiO}_2$ (a mineral of ceramic ware), alite, $3\text{CaO} \cdot \text{SiO}_2$ (a cement mineral), devitrite, $\text{Na}_2\text{O} \cdot 3\text{CaO} \cdot 6\text{SiO}_2$ and similar calcareous alkali silicates entering into the composition of industrial glasses, thomasite, $9\text{CaO} \cdot \text{P}_2\text{O}_5 \cdot \text{SiO}_2$, oldhamite, CaS , and other silicophosphates and sulphides found in metallurgical slags.

Along with the special features of their chemical composition, the temperature of formation is an important factor in the formation of the minerals of these technical aggregates. The temperatures of forming minerals are much higher than the temperatures of formation of many rocks. Due to this circumstance we find in refractories and slags, for example, that silica crystallizes as tridymite and cristobalite instead of the quartz which predominates in most natural rocks, and that pseudowollastonite occurs in the place of wollastonite, mullite instead of sillimanite, carnegieite instead of nepheline, and so on.

Systematic studies of technical rocks have revealed regular behaviour of great theoretical and practical importance. The variation in mineral composition of open-hearth slags has been traced in the course of smelting; phenomena ascribed to the liquation of a silicate melt have been revealed in molybdenum slags, and the differentiation of ferrochrome slags during smelting has been established.

A detailed study has been made of the devitrification of technical glasses in relation to composition; the problem of flaws in glass ("stones", "schlieren" and similar) has been elucidated from the point of view of the solubility of different minerals in molten glass; a parallel has been drawn with the relevant magmatic rocks (phonolites, liparites) and the products of reaction between fireclay glass-

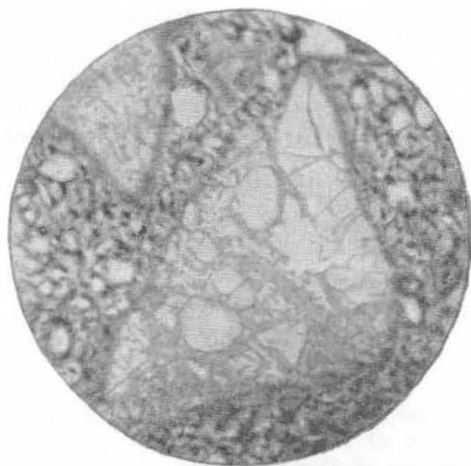


Fig. 117. Quartz fragment; cristobalite is seen in its cracks. Unaltered dinas brick

tanks with furnace dust and the alkaline vapours from the surface of the molten glass.

A mineralogical, petrographical and chemical study of different kinds of refractories during their actual use in metallurgical and glass-melting furnaces has furnished numerous examples of artificial metamorphism. The quartz brick known as dinas is especially characteristic in this respect; after long service in the lining of an open-hearth furnace, it becomes zoned and comprises an unaltered light-yellow zone, a transitional light-brown zone, a black tridymite zone, and a grey cristobalite zone (Fig. 117). The latter forms the brick capping exposed towards the inside of the furnace.

Parallel with this regeneration of quartz as tridymite, and then cristobalite which determines the zoning just described, a general chemical transformation of the zoned brick is also observed proceeding according to the so-called dinas law: an accumulation of titania and alumina in the transitional zone, and of iron oxides and manganese in the tridymite zone of the affected brick.

Petrurgy is a very interesting field of silicate technology. It involves the remelting of basalt and other rocks with additions of some sort. Its products show an improvement of mechanical and thermal properties in comparison with the original material, and they can be given any required form. The cast stone obtained in petrurgical works is a curious instance of the dependence of mechanical and thermal properties of rocks on their detailed mineral composition and structure.

CHAPTER XVII

The importance of petrography in the national economy

Mention has already been made in the preceding chapter of the prospects for a study of technical rocks by petrographical methods, and of the importance of these methods in the development of contemporary metallurgy and of the silicate products of the glass, ceramic and cement industries. The practical problems of petrography associated with the study of natural rocks can be combined into the following two large groups:

- the study of rocks genetically associated with different kinds of minerals with the aim of searching for and exploring these deposits;

- the study of rocks which are directly economic.

The first group of problems is very much wider than the second and includes all branches of petrography.

Many ore bodies are found in association with magmatic rocks, and consequently the search for them is bound up with the deciphering of magmatic processes as the primary source of such mineral concentrations.

Very wide perspectives are opened by the study of basic rocks. It is generally known that the main mass of ore deposits (non-ferrous and rare metals, gold) are connected with intrusions of acid magma, and their production has been correspondingly more studied. At the same time numerous deposits of metals and mineral raw materials connected with basic and ultrabasic rocks have been discovered in recent years. Thus, impregnations of sulphides of nickel, copper, cobalt, sometimes containing platinum metals, are found in some gabbro massifs. With other, also ore-bearing, gabbros are associated concentrations of titanomagnetite. Ultrabasic rocks, predominantly dunites, sometimes contain large-scale segregations of chromite, in some cases, with platinum. The main mass of platinum is won from placers, formed from the breakdown of bodies of ultrabasic rocks. The celebrated diamond-bearing pipes of South Africa and Yakutia are also filled with an ultrabasic rock similar to the peridotites, in which diamond plays the part of an accessory mineral.

It is difficult to cite cases of acid and intermediate rocks, in which the ore mineral might be one of the direct components of the rock; although they doubtlessly exist (for example, the tin-bearing pegmatites, and granites containing rare earth minerals), they are of very little industrial importance. Ore segregations associated with acid magmatic rocks are often concentrated at the contact zones of intrusive massifs; such are the magnetite concentrations in skarns (the Magnet and High Hills), the chalcopyrite stocks in skarns and hornfelses (the Bogoslov mining district in the Urals). The overwhelming majority of such deposits are hydrothermal segregations, most often veins or vein-like bodies which have been formed by the action of gaseous emanations and various types of solutions given off by the cooling magma.

Petrographic methods play a very important part in the complex geological investigations carried out during the search for these deposits.

Petrographic methods are no less important in studies of minerals associated with sedimentary rocks. Study of the material composition and the elucidation of the conditions of formation of any sediment are impossible without using petrographic methods. Of sedimentary rocks, limestones are of the greatest importance, being used after burning as the binding raw material or cement, and they are also one of the most important raw materials in the chemical industry, where they are used in the preparation of paper, glass, ammonia, potash, soda, sugar, and also find application in agriculture (lime fertilizers); lime is necessary for the metallurgical industry for the production of metal and the reduction of ores (500 kg of limestone is needed for the smelting of 1 ton of iron).

Wherever there are heating furnaces and baths for metal fusion, the products of the ceramic and refractory industries are required for the refractory casings with which they have to be fitted.

Concrete, asbestos-cement and ceramic ware are acquiring an ever-increasing importance as substitutes for iron and wood.

There is a very great demand for raw materials by such large undertakings as cement mills, brickworks, gravelworks and stone quarries.

The following sedimentary rocks are used in the production of building materials: (1) limestones, marbles, dolomites, marls, chalk; (2) refractory and low-melting clays, loams and slates; (3) sands and sandstones; (4) gravel and boulders; (5) tripoli earths, diatomites and gaizes; (6) gypsum, thenardite, mirabilite, etc.; (7) tuffs.

In some industries (cement, glass) the chemical composition of the rock is of the greatest importance, in others, the physical and mechanical properties, structure and texture (the production of facing materials, building stone, fillers for asphalt pastes, roofing slates and similar).

Limestone, chalk, marl, clays and slates are the basic raw materials of the cement industry; quartz sand serves as an additive, iron ores and bauxite as adjusting additives introduced into the raw mixture; gypsum, tripoli earth, diatomite and tuffs are added during the grinding of cement clinker. The waste products of the metallurgical, power supply and other branches of industry are also used in cement manufacture (slags, ash and suchlike); the composition of waste products must be taken into consideration when studying basic raw materials. CaO , SiO_2 , Al_2O_3 , Fe_2O_3 should predominate in the chemical composition of a rock which is used for the production of portland cement, because they are necessary for the reactions which take place when the clinker is fired, to form the clinker minerals. The calcium oxide is derived mainly from carbonate rocks, the silica, alumina and iron oxides from argillaceous. Detrimental substances in the raw material are magnesium oxide, phosphorus, alkalies and sulphur (especially associated with gypsum). The mineralogical analysis of argillaceous rocks is very important. The large particles of quartz, feldspars and other minerals entering into the composition of these rocks are often distinguished by a lowered capacity for reaction. No less important is a granulometric study of the raw material, especially of the argillaceous component, since the reaction capacity of a raw material is linked with the degree of its dispersion, and the relations between the coarse and fine particles in a clay are very important.

The glass industry supplies the economy with both the structural glasses proper (window, architectural) and different forms of glass with specific properties used for packaging, chemical apparatus and vessels, as well as for optical and technical appliances. Both natural rocks and minerals (quartz sand, sandstone, limestone, marble, chalk, dolomite, magnesite, feldspar, kaolin, sodium sulphate, witherite, soda, fluorite) and also artificial minerals (potash, industrial alumina, boric acid, etc.) are used in glass production. Various amounts of the starting materials are used in making up charges for glass melts of one composition or another, the basic role belonging to quartz sands, and to a lesser degree to the carbonate rocks. The purity and homogeneity of the starting material are deciding factors which determine the quality of the glassware, and affect the course of the technological process. The requirements of industry with regard to glass sand are quite rigorous, and are defined as a content in the sand for window glass of not more than 0.2% of iron oxides, and not more than 0.02% of chromic oxides. These requirements are still higher for special varieties of glass. Therefore, natural sands answering these conditions are very sparse, and the reserves of the deposits which are known are as a rule extremely limited. Only isolated deposits of quartz sand have reserves of a few millions of tons.

The following mineral raw materials are used in the structural ceramic industry: refractory, high-melting and easily fused clays

and kaolin (as the basic components of the working ceramic mass), quartz sand, feldspar, pegmatite, limestone and some other economic mineral materials.

In the industry of structural cementing materials, and in the production of structural components, gypsum is of great importance; it is widely used for the production of dry gypsum plaster, gypsum blocks and architectural details.

Limestones, dolomites, marbles and also sandstones and quartzites are used in industry as natural facing materials. Rocks used as raw material for the production of facing stone should have a high decorative value (beautiful colouring or design), be sufficiently free of joints to permit the production of blocks of sufficient bulk, and be resistant to mechanical and meteorological actions.

We shall now examine the properties and uses of the individual varieties of sedimentary rocks.

Clays. Clays contaminated with sand are used in building, and as material for making different sorts of brick; the purer ferruginous varieties are for pottery; the pure kaolin clays serve as the raw material for the production of porcelain and faience. Fulling earth, used for the removal of coloured organic materials, and clays capable of selective absorption of high-molecular vegetable and animal fats and paraffins consist predominantly of the mineral montmorillonite. They are used for the purification of petroleum, nutritive oils and fats, and also for the removal of grease from wool. Kaolin clays are used as a filler in the paper and rubber industries, and also in soap making. The coloured varieties of clay can be used as mineral colours (ochres and umbers). Clay also serves as the raw material for the production of aluminium compounds, alum and ultramarine.

Sands. This is one of the commonest stony structural materials. The hardness, resistance to acids, refractoriness, chemical affinity for alkalies permit quartz sand to be used in a number of branches of industry. Quartz sands are used as abrasives; a contaminated variety is used as ballast, a concrete filler, a filler for water filters, and for the preparation of silicate brick. Exceptionally pure quartz sands are used as the raw material in glass production, the most deleterious minerals being those of chromium and iron (they give a green tint to the glass), and also the high melting minerals (zircon, kyanite), whose presence causes flaws in the glassware. Sands of a definite coarseness, free of easily fused material, can be used as a moulding material.

Limestones. Pure limestones are used as a raw material in the preparation of glass; the calcination and slaking of lime is used in the preparation of structural cementing solutions; in metallurgy, limestone is used as a flux; crushed and calcined limestone is an excellent fertilizer for soils poor in lime. Limestones are used in the

preparation of bleaching powder, which is used in textile and paper mills, and in the preparation of soda ash, which enters into soap making and is widely used in the manufacture of glass, chemicals and textiles. Marly limestones serve as the raw material for portland cement.

Marble is used for internal and external facings of buildings, as a decorative, sculptural and ornamental stone. The way in which a marble is utilized is determined by its granularity, colour, and the dimensions of possible monoliths (free of joints or cracks).

Gypsums find application in the cement industry, in architecture, in agriculture for fertilizer production, in the chemical industry for the preparation of ammonium sulphate and sulphuric acid; the main application of gypsum is in the preparation of plaster of Paris; the purer the gypsum, the better the adherence and strength of the plaster.

The acid refractory known as dinas is prepared from *quartzites* and *sandstones* free of impurities ($\text{SiO}_2 > 97\%$), with a small addition of lime as a binding agent.

Dunites can be used for the preparation of a highly refractory material suitable for the vaults of open-hearth and electric steel melting furnaces. Due to their high magnesia content (up to 50%), they can be used as an ore of metallic magnesium, and as a raw material in the production of fertilizers. Serpentine is also used in this way.

If *nepheline rocks* do not contain any iron minerals, they are a valuable raw material for the glass and ceramic industries; they can also be used for the fusion of titanomagnetites as a flux, and when large amounts of cheap and pure nepheline are available, it is a good ore for smelting aluminium.

A method of obtaining a crystalline product from fused *basalt* has been discovered comparatively recently. A basalt casting possesses high insulating properties, resistance to acids, and impermeability to water, as well as a great resistance to crushing (above 3700 kg/cm^2). Many kinds of wares are made out of fused basalt: acid-proof components, pipes, insulators, floor tiles, balls for grinders, and so forth. The production of basaltic padding is closely related to the production of fused basalt, and is used in construction work.

Mineralogical and petrographical investigations are of great importance in ore concentration and in the working of other types of mineral products. They give an idea of the composition of the ore mass, and make it possible to give an answer to such questions as: is it necessary to break up the concentrates from the first treatment; what is the character of losses of some minerals and metals; what are the impurities found in the concentrates; must they be got rid of, or is there a possibility of their incidental separation and utilization. Such investigations have been successful in revealing new forms of aluminium and lithium ores, and pointing the

way to the interdependent utilization of all the material obtained from the working of a deposit (including coal deposits) which had previously been considered uneconomic.

The utilization of rocks as economic materials requires specific petrographic investigations, and investigations of the physical and mechanical properties of the rocks.

Petrographic investigations during the opening up stages of mining operations can determine any resistance to boring that the rocks may have, and form a study which impinges directly on this branch of petrography, being connected in the closest way with the petrographic and physical and mechanical properties of the rocks.

CHAPTER XVIII

The elements of engineering petrography

1. THE CONCEPT OF ENGINEERING PETROGRAPHY

The engineering requirements for rocks are mainly concerned with their mechanical and chemical properties; engineering petrography is the study of these features by petrographic methods. This branch of petrography is connected with mining problems (rock pressure, mechanical and thermal drilling) as well as with problems of the building stone industry and of chemical raw materials, and with several other offshoots of the national economy.

2. THE PHYSICAL AND MECHANICAL PROPERTIES OF CRYSTALS

The physical and mechanical properties of crystals include those related to such mechanical effects as impact, compression, tension, and so on. These properties, like others, are controlled by the structure of specific crystals. This is shown especially clearly in their cleavage and plasticity.

Mechanical forces applied externally to a crystal give rise to deformations, which can be plastic or elastic. A deformation is said to be plastic if the change in form is preserved after the action of the deforming forces has ceased, and elastic if the body resumes its original form after the forces have been removed.

Examples of plastic deformation are gliding and mechanical twinning.

Gliding is a parallel change of position of layers in a crystal along a constant (stationary) plane, which is called the glide plane. The magnitude of displacement of the individual layers is directly proportional to their distance from the glide plane. The portions of the crystal retain their mutually parallel orientation during the gliding.

In *mechanical twinning*, there is likewise a displacement of layers along the twinning plane, but in this case the displacement is subject

to some twin law. The parts of the crystal change their relative orientation.

By *hardness* is implied the resistance to penetration by the point of another body. The hardness of a mineral is usually evaluated on a ten-point system by scratching it with another mineral, adopted as a standard (Mohs' scale). Hardness is determined more accurately by special appliances called sclerometers. There are a great number of different sclerometers, which sometimes evaluate this property quite differently. Thus, with the Martens sclerometer, the hardness is determined by the load in grammes pressing on a diamond point, which leaves a scratch of the set width (0.01 mm) on the given body.

Besides scratching, there are other methods of testing hardness. The abrasion, for example, which is obtained by the impressions of a steel ball under pressure with a given load, and so on. A microscope appliance constructed by Khrushchev and Berkovich has recently come into widespread use. It is a diamond point pressing with a standard load on to the test material. The size of the impression obtained is the measure of the hardness.

Hardness has been shown to depend on the structure, the distance between the structural units, and on the valency of the chemical constituents. For the same structure and valency, the hardness increases with decrease in the radii of atoms or ions forming the crystal lattice. For similar types of structure and equal or nearly equal distances between the ions (the sum of the radii of adjacent ions), the hardness increases with valency. The high hardness of diamond is explained by the conjunction of the high valency of carbon (4) with the small distance between the centres of its atoms (1.54 Å).

The hardness of different faces of the same crystal can be dissimilar when the packing density is different on the different faces.

It is easier to scratch a face along rows of closely arranged particles and more difficult across them. The mineral disthen or kyanite, Al_2SiO_5 , is an example of this. Its crystals are elongated tablets thickened parallel to the face (100). A steel knife scratches this face easily across the elongation of the crystal, but will not scratch it along its length.

In *elastic deformation*, after removal of the deforming forces, the crystals spontaneously, under the influence of internal forces, return to their original state. The alteration preserved after the cessation of mechanical action on the crystal is called the residual deformation.

Cohesion forces. All crystals without exception are anisotropic bodies, and consequently their elastic limits are different in different directions in the general case in all crystals. The number of equal directions in relation to the elasticity of the crystals can prove to be greater than the corresponding symmetry of the crystals in question. In these cases, we speak of the cohesive forces of the crys-

tal. Besides this, the elasticity curves possess an inversion centre even when it is absent in the geometry of the crystal faces.

The thermal conductivity of crystals is expressed by the surface of a sphere for crystals of the cubic system, the surface of an ellipsoid of rotation for crystals of intermediate systems, and by the surface of a triaxial ellipsoid for crystals of the lower symmetry systems.

The distribution of crystals according to thermal conductivity is analogous to their distribution according to optical properties, the orientation of the ellipsoid of thermal conductivity being the same as for the optical indicatrix.

3. THE PHYSICAL AND MECHANICAL PROPERTIES OF ROCKS

The mechanical properties of rocks depend on:

(1) the qualitative and quantitative mineral composition of the rock;

(2) the structure (granulometric composition) and texture of the rock, the internal structure of the crystals, the character and properties of the intercrystalline cement.

The ultimate strength of a rock is determined by testing samples to destruction under compression, tension, flexure and shear.

Many sandstones which, in spite of a high quartz content, have negligible strength, will serve as examples of the dependence of the strength of a rock on the bonds between the mineral grains. Very jagged fine grains give a higher strength than a sandstone with rolled grains of unequal size. The concept of "strength" is not a simple one, since it refers similarly to a resistance to pressure, to tension and to impact; and at the same time, the demands made on the rocks differ in various cases. Therefore, the future use of a rock should be determined precisely before it is tested. Very great differences are observed, for example, between the resistance to pressure and the resistance to impact in compact rocks. Exactly the same two rocks with the same resistance to crushing can offer different resistances to attrition. The resistance of a rock to fracture is approximately 1/10-1/30 of the resistance to crushing. Some physical and mechanical properties of rocks important in technology are given in Table 14.

Data for the most important physical and mechanical properties of rocks are given in more detail below.

Resistance to compression (to crushing). The variation limits of the strength of the most important building stones are given in Table 15.

The resistance to compression is determined on samples shaped as a cube. The sample is crushed on a powerful press, and the greatest value obtained for the strength is referred to the unit area of the sample.

Table 14

Some Physical and Mechanical Properties of Rocks (average values)

Rock	Bulk weight of rock, g/cm ³	Specific gravity, g/cm ³	Moisture capacity, % (by wt.)	Resistance to crushing (in dry state), kg/cm ²	Strength of chippings (30-60 mm) on impact (degree of destruction)	Modulus of elasticity, kg/cm ²	Bulk weight of chippings (30-60 mm), ton/m ³
Granite	2.60-2.64	2.62-2.85	0-0.5	1600-2400	0.7-1.1	500-600	1.31-1.40
Porphyry	2.55-2.64	2.58-2.66	0.2-0.5	1800-3000	0.6-1.3	500-700	1.30-1.36
Diabase	2.80-2.86	2.85-2.95	0.2-0.4	1700-2500	0.5-0.9	700-800	1.36-1.45
Basalt	2.95-3.00	3.00-3.15	0.1-0.3	3000-4000	0.5-1.0	900-1200	1.40-1.50
Vesicular basalt lava	2.20-2.35	3.00-3.15	4.0-10	800-1500	—	400-500	1.10-1.25
Quartzite	2.70-2.68	2.69-2.72	0.2-0.5	1500-3000	0.8-1.4	400-600	1.25-1.35
Quartzitic sandstone	2.60-2.65	2.65-2.70	0.2-0.5	1200-2000	—	300-400	—
Other sandstones	2.0-2.60	2.64-2.72	0.2-9.0	300-1800	—	50-300	1.25-1.35
Ordinary limestones and dolomites	1.70-2.85	2.70-2.90	0.2-0.5	800-1800	0.9-1.6	400-700	1.30-1.35
Ditto, weak	—	—	0.2-10	200-900	—	300-600	—
Gneiss	2.65-3.00	2.67-3.05	0.1-0.6	1600-2800	0.7-1.2	—	1.30-1.50
Amphibolite	2.70-3.10	2.75-3.15	0.1-0.4	1700-2800	0.5-0.8	—	1.40-1.50

Table 15

Breaking Point on Compression or Crushing, kg/cm²

Rocks	Very strong	Strong	Medium	Weak
Granites	2300-3700	1600-2200	1000-1500	600-900
Porphyries	2500-3500	1800-2400	1200-1700	500-1100
Basalts	3500-4500	2800-3400	1300-2400	800-1200
Sandstones	1500-2200 ¹	900-1400	500-800	250-400
Limestones	2000-2500	1200-1900	800-1000	200-700 ¹
Marbles	1500-2000	1200-1400	900-1100	300-800

Porosity. The direct determination of porosity comprises a comparison of the specific gravity of a piece of the rock, δ , with the specific gravity of its powder, γ . The porosity coefficient is

$$P = \frac{(\gamma - \delta) 100}{\gamma}$$

The porosity can be estimated indirectly by the absorption of water, determined by comparing the weight of a broken-off portion of rock dried at 50°C, and the same piece after maximum imbibition of water.

The coefficient of water saturation S indicates the ratio between the weight of water W_1 imbibed at ordinary pressure and slow loading, and the weight of water W_2 under a pressure of 15 atm, i.e.,

$$S = \frac{W_1}{W_2}.$$

It is important that the saturation coefficient does not exceed 0.8, if the resistance to frost is to be determined. If the pores are more than nine tenths full of water, the water, on freezing, will exert a pressure on the walls of the cavities and shatter the rock. The limiting value of this coefficient is therefore taken as equal to 0.8.

Frost resistance is determined by freezing 25 times in a special apparatus where temperatures down to -35°C can be attained.

Softening in water. The coefficient of softening is

$$n = \frac{K_{wet}}{K_{dry}}$$

where K_{wet} = resistance to compression of the rocks when water-saturated

K_{dry} = the same, in the dry state.

Crystalline rocks do not soften at all or only slightly. Rocks with a greater or smaller content of clay soften very much.

Table 16

Attrition of Rocks on a Standard Circle, mm

Rock	Very hard	Hard	Medium	Soft
Granite	0.9-1.1	1.2-1.4	1.6-1.8	2.2-2.6
Porphyry	0.7-0.9	1.0-1.2	1.4-1.6	1.8-2.4
Basalt	0.8-1.0	1.1-1.3	1.5-1.7	1.9-2.9
Quartzite	0.5-0.7	0.8-1.2	1.3-1.6	1.7-2.0
Sandstone	1.4-3.0	4.0-5.0	6.0-7.0	8.0-12.0
Limestone	2.8-4.0	5.0-6.0	7.0-8.0	9.0-11.0
Silicified limestone	1.2-1.5	1.7-2.4	2.8-3.2	3.6-4.0

Wear (abrasion or attrition) of rocks is determined by polishing them with emery cloth on a horizontal rotating iron plate. The wear of the sample (mm) serves as a measure of the abrasion during mechanical polishing under standard conditions (Table 16).

Climatic durability. Crystalline rocks are more stable than glassy or amorphous ones, the fine-grained being of a greater durability than the coarse-grained. A uniform grain is important for climatic stability, as well as the absence of pores and cracks and ingredients such as pyrite or bituminous material which is easily converted into an acid.

Resistance of rocks to fire is tested by heating a lining of the test rock in special furnaces and then pouring water on to them.

Durability. The mechanical properties do not give the structural characteristics of a stone directly, but a number of these characteristics can be figured out by making use of the mechanical properties in conjunction with a consideration of other petrographic features of the stone. Data concerning the theoretical longevity of a stone can be cited as an illustration. This rock characteristic is obtained by comparing the mechanical strength of the rock, its resistance to freezing, destruction by water and similar properties

Table 17

Theoretical Durability of Rocks (in years)

Rock	Start of disruption	Imminent state	Final disruption
Granites	75-350	220-1070	650 to 1625
Quartzites	220-475	675-1460	> 1625
Marbles	20-135	33-400	100 to 1200

Table 18

**Theoretical Durability of Carbonate Rocks
of the Samarskaya Luka, U.S.S.R. (in years)**

Rock	Underwater dam structures			Above high-water level			In zone of changing levels		
	start of disruption	imminent state	final disruption	start of disruption	imminent state	final disruption	start of disruption	imminent state	final disruption
Dolomite	220	370	900	220	450	—	20	42	105
Calcareous dolomite	220	370	900	220	450	—	20	42	105
Dolomitized limestone	30	65	160	38	80	195	—	22	50

that are obtainable from laboratory tests of rock samples, selected in such a way as to reproduce the conditions in which the rocks will fulfil their function as parts of a construction in an urban climate. Some results of such tests are given in Tables 17 and 18.

4. THE DEPENDENCE OF THE ENGINEERING PROPERTIES OF ROCKS ON THEIR PETROGRAPHIC FEATURES

The dependence of the properties of a rock on structure, mineral composition and preservation (weathering) can be ascertained from a study of a thin section of the rock under the microscope.

Igneous Rocks

Medium- and fine-grained rocks possess a higher strength than coarse-grained. Rocks which have been subjected to tectonic action, and jointed rocks take up water; therefore, they are not frost-resistant and disintegrate in construction work. An abundance of mica lowers the strength. A porphyritic structure is advantageous, especially when inclusions are few and fresh, and the groundmass is finely crystalline and continuous (without pores and cavities). Large inclusions with cracks lower the strength, as also does a groundmass rich in glass. Rocks with a felted arrangement are very tough and difficult to bore. These properties are as follows in individual rocks:

Granites. Favourable signs: freshness of feldspars, an insignificant mica content, high quartz content, an absence of pyrites, and an equigranular finely crystalline texture.

Gabbro and gabbro diabases. Favourable signs: freshness of feldspars, a high content of hornblende or pyroxene, an absence of serpentinization or chloritization of these ingredients, an absence of pyrites.

Basalts. Favourable signs: the predominance of augite and a low content of olivine, an absence or insignificant amount of a glassy base, microcrystalline texture, dark colour, smooth fracture surface. Unfavourable signs: the predominance of feldspar, a large olivine content, the presence of nepheline, non-uniformity of grain size, grey colour, white stellate patches (sunburn).

Andesites and trachytes. Approximately the same signs. Any considerable amount of mica must also be avoided, and preference given to a high content of hornblende or pyroxene; porosity is a contraindication in trachytes.

Quartz porphyries. Favourable signs: a high quartz content, a small amount of mica, an absence of pyrites, freshness of feldspars, a microcrystalline or aphanitic texture, smooth or splintery fracture, fresh when fractured.

Sedimentary Rocks

The majority of stratified rocks are distinctly anisotropic; they usually have different properties in directions parallel and perpendicular to the bedding. This situation can be illustrated by Table 19

Table 19

**Physical and Mechanical Properties of Typical Sedimentary
Rocks of the Donbass**

Rocks	Elastic modulus $E/10^3$, kg/cm ²		Poisson's ratio, μ		Ultimate strength, kg/cm ²	
	on compression					
	to the bedding	⊥ to the bedding	to the bedding	⊥ to the bedding	to the bedding	⊥ to the bedding
Slates	316	154	0.22	0.22	367	517
Sandy and clay shales	124	44	0.20	0.08	172	266
Sandy shales	303	242	0.25	0.16	518	789
Limestones	636	725	0.28	0.30	1510	1250
Coals	—	—	—	—	61	69
Anthracites	42	54	—	—	105	160
Sandstones	347	398	0.13	0.13	841	1920

which gives the ultimate strength, elastic modulus and Poisson's ratio for some typical rocks from the Donbass.

The mechanical performance of rocks on bending is of great practical importance, since rocks are subjected to deformation in the form of flexure as a result of underground working. In the first period, the bending is the consequence of different crustal movements, in the second, it arises as a result of the redistribution of stresses around the mining development and the squeezing out of the lateral rocks into the worked-out spaces.

Limestones. Favourable signs: an equigranular, fine or medium crystalline texture, dense structure (the absence of porosity can be easily confirmed, if a drop of ink is washed off after a few minutes and does not leave a noticeable stain on the polished surface), an absence of pyrites, especially of the finely disseminated kind which gives the rock a grey or bluish colour, an absence of intercalations of serpentine, mica and similar minerals, a more or less reasonable thickness of bedding, a coarse jointing, an absence of finely bedded stratification and of irregular fortuitous cracks.

For compact dolomites and limestones, indications are: a micro-crystalline or massive structure; freshness and a white, yellowish, greyish, greenish or reddish colour and an absence of pyrites; a more or less shining, fine-grained fracture surface, considerable hardness (when tested), an absence of porosity revealed by a negligible change in hardness and resistance to compression in a water-saturated condition; an absence of an argillaceous smell; fracture and stratification as in marble. Sometimes porous and oolitic or massive limestones rich in fossils prove to be fairly stable, if they have a fine-grained texture.

Sandstones. In sandstones suitable for building, the grains should be exclusively, or almost exclusively, quartz, with only a negligible amount of isolated grains of fresh feldspar, and sometimes a little glauconite and mica; quartz cement is better than carbonate or argillaceous. If the cement between the grains is quartzose, the cement filling the pores can be argillaceous, carbonate or similar. Sandstones with a carbonate cement are unsuitable for underwater constructions, or for constructions which are subjected to the action of water or smoke. A mullion structure (an intergrowth of grains in direct contact with each other) endows sandstones with good physical and mechanical properties.

Crystalline Schists

Crystalline schists mostly reveal still sharper differences in strength in directions parallel and perpendicular to the schistosity than do the sedimentary rocks. Only completely fresh and firmly coherent varieties of schists find a technical use. The arrangement of the mica flakes is critical for roofing slates.

Gneisses. Favourable signs are the same as for granite. The most favourable rocks are those in which the gneissose character is not very strongly developed and which do not possess a very fine jointing or smooth flat bedding planes.

Roofing slate. Favourable signs: dark greyish or greyish-black colour, good fissility into fine slabs which can be pierced by a nail without breaking, a silvery sheen on the foliation planes; considerable hardness and a sharp sound when struck are evidence of favourable features of structure and composition, while these latter features are expressed in a large content of mica, which should be arranged in the form of connected, sometimes wave-like, winding bands, in the absence of pyrites and sometimes in the high content of silica. A dull black colour is a contraindication since it shows a considerable content of bituminous material.

General Evaluation of Rock Quality

An obligatory quality of rocks destined for technical use is freshness, i.e., a good state of preservation, the minerals building up

Table 20

Approximate Evaluation of Rock Quality from External Features

Feature	Features of high-quality rocks	Features of rocks which are weak or unsuitable
Sound when split by hammer Smell Fracture	Clear, pure tone Absent Smooth, conchoidal, shining crystalline surface; no cracks; sharp edges	Dull, rattling sound Clayey smell Flocky to tubercular; fracture surface with iron oxide films; cracks; edges blunt and crumbling
Size of chip	Possible to chip off a large piece with fresh fractures without cracks (estimated in m)	Large pieces are not chipped off, the piece is bounded by old cracks
Weathered crust along cracks	Absent or a thin film	A wide weathered crust; films of iron oxides
Grain size	The rock is fine-grained to dense	The rock is coarse-grained; the minerals fall out

the rock being in an unweathered condition, and also the absence of jointing. The quality of the rock can be judged from known external features (Table 20).

5. THE EFFECT OF GEOLOGICAL PROCESSES ON THE QUALITY OF BUILDING STONE

The effect of geological processes on the character of a rock is expressed to a considerable extent in its quality as a building stone. Thus under the influence of dynamic processes (such as tectonic dislocations), rock frequently reveals an internal crushing of its ingredients which is indiscernible to the unaided eye, and an increasing brittleness which creates a system of microjointing, which in its turn increases the moisture capacity of the rock and greatly decreases its frost resistance. The schistosity of some varieties of rock is due to the same causes, which destroy the physical uniformity of the rock and substantially influence the value of its breaking point.

The degree of metamorphism is of great importance in estimating the building quality of a rock. The conversion, for example, of sandstones into strong and compact quartzites is due to metamorphism. Normal diabases are converted by the same processes into hornblende diabases, thereby, on the one hand, increasing the tenacity of the rock, and on the other, increasing the difficulty of its working.

The importance of metamorphism is especially graphically revealed in the case of limestones and dolomites, which, on the one hand, take on a homogeneous crystalline granular texture, and on the other, are enriched in quartz. The actual dolomitization of limestones is a phenomenon of the same order.

As an example of a metamorphic quartzite, let us consider the compact crimson quartzitic sandstone of Shoksha in the U.S.S.R. Although this rock is distinguished by a considerable hardness, due to its great uniformity of structure, it is easily broken up by wedges and gives a good and even fracture. The material lends itself to working and polishing. Due to its high technical quality and its pleasant uniform tone, Shoksha sandstone has long been used for a succession of works of art. In particular, it was used in the execution of many of the details of Lenin's Mausoleum.

The hardness of the Shoksha quartzite almost reaches the hardness of quartz, and it is therefore characterized after polishing by a perfect brilliancy. Its technical quality is extremely high. The breaking point on compression is 1729-1894 kg/cm² in the saturated form, and 1712-2218 kg/cm² in the dry. It withstands a 25-fold freezing, without suffering any noticeable change.

6. DRILLING FEATURES OF ROCKS

The basic method of disintegrating rock in the mining industry when working minerals is by drilling bore and blast holes, with subsequent blasting of the rocks.

Mechanical Drilling

Rocks consist of minerals with different physical and mechanical properties, including different strengths. During drilling, any one of the constituent minerals is broken down to extraordinarily fine fractions, so that all the minerals suffer equal destruction. The resistance of all minerals must therefore be considered when drilling. Coarse-grained rocks, for example, are always less strong than fine-grained rocks of the same composition.

The character of the cementing material, as well as that of the grains, has a substantial influence on the mechanical properties of detrital rocks. If the mechanical strength of the grains in the rock is higher than that of the cement, the fracture takes place in the cement, in the opposite case, in the grains; if the strength is the same in both, the fracture takes place partly in the cement, partly across the grains.

Porosity is the next main factor affecting rock strength. The existence of porosity creates a weakened region in the rock and aids the penetration of the drill, but on the other hand, it gives rise to chemical reactions between the rocks and the surrounding medium when water orgases are permeating the rock. Water weakens the bonds between the particles of the rock and leads to a diminution in strength. Data obtained by B. Zaleskii and others corroborate the connection between porosity and strength (the breaking point on compression of rocks).

Tables 21 to 25 give drilling data obtained by Zaleskii for several rocks.

The following relationships can be said to hold within individual groups of rocks:

1. The drilling capacity of limestones depends on texture, composition and degree of freshness and is absolutely unconnected with mechanical indications, particularly the attrition.
2. The drilling capacity of Devonian sandstones depends both on the petrography and on the strength (attrition and crushing).
3. A similar relationship is not observed in jaspilites.
4. The drilling capacity of quartz-feldspar rocks depends on the petrographic composition, the strength and the hardness.
5. For jasper, an especially compact rock, rich in free silica, the dependence of drilling capacity on the resistance to attrition and crushing is analogous to that of the quartz-feldspar rocks.
6. For nephrite, an especially tough rock, there is a characteristic absence of any connection between its drilling capacity and

attrition and hardness, in comparison with quartz feldspar rocks. The rate of progress is very low.

An investigation of the secondary quartzites of Kounrad, conducted by the author in conjunction with A. Medvedko and S. Ivanovskii, has shown that the rate of drilling through them under similar conditions is directly proportional to the degree of sericitization of the secondary quartzites.

Thermal Drilling

The so-called thermal drilling of bores and blast holes differs in principle from mechanical drilling and is characterized by an increased rate of drilling, a low consumption of explosives and a small number of blast holes on the work face. All known methods of thermal drilling are divided into two types, according to their action on the rock: (1) those based on thermal disruption (exfoliation method); (2) those based on rock fusion.

In the first case, the heat source is a stream of gases from special reaction burners, which is ejected on to the rock face at great speed (about 2000 m/sec) and at a high temperature. The high temperature heats up a thin layer of rock on the work face, which expands and splinters off, stripping the surface. The rock splintered off is shattered by the stream of gases, and the fragments are borne out of the drilling before they have time to melt.

In the second case, the source of heat (exothermic combustion of iron in oxygen—oxygen blast, heat from an electric arc, and so on) is concentrated on the surface of the wall of the drilling, and gradually

Table 21

Lower Carboniferous Limestones

Type	Texture	Composition, %		Average penetration with hard alloys, mm/min	Attrition, g cm ²
		crystalline granular portion	pelitic-morphic mass (cement)		
Organic-detrital	Crystalline granular	70	30	8.5	0.575
Ditto	Irregularly grained detrital	40	60	48	0.560
Ditto	Ditto	50	50	54	—
Detrital	Detrital-coagulated	35	65	63	0.700
Ditto	Ditto	25	75	66	—

Devonian Quartz Sandstones
(average grain size 0.1-0.5 mm)

Table 22

Type	Texture	Cement content, %	State of preservation	Rate of progress of crusher, mm/min	Attrition, g/cm ³	Pressure breaking point, kg/cm ²
Compact quartzitic sandstones (7 specimens)	Granular and regenerated	3-7	Fresh	2.5-5	0.10 { 0.08 0.11	1300 { 1800 800
Medium- and coarse-grained sandstones, weakly kaolinized or ferruginized (6 specimens)	Granular with film and cortical cementation	10-15	More or less fresh and little decomposed	12-31	0.15 { 0.13 0.19	500 { 650 400
Kaolinized, irregularly grained sandstones (4 specimens)	Irregularly grained with crystallized cementation	10-20	Decomposed	41-85	0.23 { 0.20 0.26	400 { 550 275

Quartz-Feldspar and Feldspar Rocks

Table 23

Rock	Mineral composition, %			Grain size, mm	Rate of progress of crusher for 30 min. in	Attrition, g/cm ³	Crushing, kg/cm ²	Impact, kg/cm ³	Hardness (Rostwal)
	quartz	feldspar	coloured components						
Granite	46	49	5	0.5	0.120	0.15	2900	66	7358
Gneiss	28	46	26	0.37	0.130	0.25	2500	102	5183
Diabase	2	49	49	0.42	0.160	0.29	3500	360	2645
Gabbro-diorite	—	34	66	0.59	0.240	0.45	2400	91	1660
Amphibole-biotite schist	—	—	—	0.45	0.440	1.23	640	16	—

Jaspilites

Type and structure	Composition, %		Progress of crusher, mm/min	Attrition, g/cm ³
	quartz	ore material		
Finely bedded ferruginous quartzites	45	50	0.75	0.049
	45	55	0.77	0.065
Banded, sometimes jointed ferruginous quartzites	70	30	0.61	0.042
			0.75	0.059
			1.20	—
			1.35	—

heats up the rock, melting it in situ, and the melt is then removed to the outside.

In both cases, it is important to have a good knowledge of the qualitative and quantitative mineral composition of the rocks through which the drilling is proceeding, and also of their chemical composition, structure and texture, microtectonics, i.e., a knowledge of the petrographic features of the rocks in order to consider their effect on the drilling process.

A rock is an aggregate of minerals which experience non-uniform thermal expansion. Mechanical stresses (tension, compression, shear) therefore arise between the rock's ingredients. This results in the formation of cracks, exfoliation and the disturbance of the continuity of the rocks. The disruption of a rock by thermal action is thus controlled by the physical properties of its constituent minerals and the thermal properties of the rock as a whole (heat capacity, thermal conductivity, and so on) (Table 26).

In crystals of intermediate and lower symmetry, thermal expansion proceeds along an ellipsoid, in cubic crystals along a sphere.

Table 25

Extra-Compact and Extra-Tough Rocks

Rock	Rate of progress of crusher for 30 min, m	Attrition, g/cm ³	Compression, kg/cm ²	Impact, kgcm/cm ³	Hardness (Rostwal)
Jasper	0.015	0.06	3850	18	—
Nephrite	0.020	0.16	5500	200	1500

Table 26

Thermal Characteristics of Rocks

Minerals and rock	Density, kg/m	Thermal conductivity, Cal/m hr deg	Temperature conductivity ($\alpha \cdot 10^3$), m hr	Heat capacity, Cal/kg·deg
Quartz	2600	—	—	0.19
Granite	—	2.06	—	0.19
Porphyry	—	1.87	—	—
Basalt	—	2.02	—	—
Serpentine	2670	3.16	0.346	0.322
Siliceous sandstone	2630	1.72	2.85	0.230
Chlorite schist	2760	3.14	3.30	0.344
Limestone (98% CaCO_3)	2480	0.846	2.65	0.129
Chalk	2400	0.80	1.91	0.21
Marble	2700	1.12	1.98	0.10
Alumina	1845	0.80	1.85	0.26

When calcite is heated, it expands in one direction, contracting in the remainder; the expansion of quartz is twice as great in one direction as in others.

When a crystal is heated, the distribution of temperature is not uniform. If the temperature on a heated surface of quartz is 60°C , it will be 5°C lower at a distance of 1 cm from the surface. This gives rise to a non-uniform expansion of the crystal and can lead to its disruption.

Polymorphic transformations of minerals take place with a change in volume. This is especially important when investigating quartz-bearing rocks (Table 27).

Table 27

Thermal Characteristics of Different Modifications of Silica

Polymorphic transformations	Volume change, %	Transformation temperature, $^\circ\text{C}$
β -quartz \rightarrow α -quartz	+ 2.4	573
α -quartz \rightarrow α -tridymite	+12.7	870
α -quartz \rightarrow α -cristobalite	+17.4	1200-1350
α -tridymite \rightarrow α -cristobalite	+ 4.7	1470
α -cristobalite \rightarrow melt	+ 0.1	1713
β -cristobalite \rightarrow α -cristobalite	+ 5.6	180-270
γ -tridymite \rightarrow β -tridymite	+ 0.6	117
Molten quartz \rightarrow α -cristobalite	+ 0.4	~ 1200

The information presented in this book is far from complete, but it shows that present-day petrography is of great practical importance. Its methods and data are utilized in the most diverse fields of contemporary technology.

In the Soviet Union, there are numerous research institutes and specialists working on geological expeditions, in mines and works whose concern is with questions of practical petrography (including technical and engineering petrography) and with theoretical problems connected with this science.

The great work of making a geological study of the country's wealth has been carried out in the U.S.S.R. Many hundreds of new deposits of minerals have been revealed and absorbed into the national economy. In this work, petrographers have taken an active part.

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